# ARCHITECTURE OF THE UPPER SEGO SANDSTONE, BOOK CLIFFS, UTAH

A Thesis

by

STANLEY SCOTT BIRKHEAD

Submitted to the Office of Graduate Studies of
Texas A&M University
in partial fulfillment of the requirements for the degree of

MASTER OF SCIENCE

December 2005

Major Subject: Geology

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Approved by:

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| Chair of Committee, | Brian Willis |
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| | Beth Mullenbach |
| | John Spang |
| Head of Department, | Richard Carlson |

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# ABSTRACT


Architecture of the Upper Sego Sandstone, Book Cliffs, Utah. (December 2005)

Stanley Scott Birkhead, B.A., Texas A&M University

Chair of Advisory Committee: Dr. Brian Willis

This study maps the facies architecture and geometry of stratigraphic surfaces within the tide-influenced upper Sego Sandstone exposed in the Book Cliffs between Crescent Junction and Thompson Springs, Utah. A bedding diagram was constructed by correlating 32 measured stratigraphic logs across this 8.5 kilometer strike-oblique outcrop to interpret depositional environments and the sequence stratigraphic setting of this tidally-influenced sandstone. Five facies associations are defined: marine shale, lower shoreface, tidally-influenced bedsets, heterolithic tidal bedsets, and tidal flat deposits. Vertical facies trends define two sandy intervals separated by a marine shale, that are interpreted to record episodic progradation of deltaic shorelines. Erosion surfaces at the base of these intervals are interrupted to record tidal scouring of the sea floor during falling stage regression. Sandstone-bodies within these intervals shingle westward recording delta lobes that thinned and became more heterolithic. Although sandstone intervals are interpreted to record progradation, internal cross stratification is dominantly tidal-flood oriented. This is interpreted to record preferential preservation of bedload transported by flood tidal currents onshore, even though net sediment was directed

offshore in a suspended ebb-oriented hypopycnal plume and as fluid mud during uncommon river floods. Deposits above high-relief erosion surfaces observed to cut down into the upper Sego Sandstone do not meet the criteria for incised valley fills. These surfaces are interpreted to record tidal current enlargement of distributary channels after abandonment. Such incisions thus do not necessarily record changes in sea level.

## ACKNOWLEDGMENTS

The idea for this research follows Dr. Brian Willis' continuing study of the Sego Sandstone. Some funding for this study was provided from Texas A&M University. Dr. Brian Willis, Dr. Beth Mullenbach, Dr. John Spang, and Dr. Michael Heaney, provided advice, experience, and friendship. Stephanie Walkup-Birkhead is thanked for her tremendous help as a volunteer editor of the manuscript. Another study, running concurrently with this research, is being conducted by Eric Robinson, another graduate student at Texas A&M University. In the Westwater Canyon area of the Book Cliffs, Matt Fey is conducting a similar survey of the lower Sego Sandstone lithofacies. Thanks are due to these two for assistance in data collection and for the numerous conversations regarding comparisons of our respective areas. Frank Bain from the Bureau of Land Management in Moab, Utah added significantly to this study by his willingness and interest in adding to my geologic knowledge of the Book Cliffs.

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# INTRODUCTION

Fluvial-, wave- and tide-dominated delta deposits have distinct vertical facies log trends, lateral and vertical facies correlation patterns, and axial-lateral and proximal-distal sand trends (Reading, 1982). Fluvial-dominated deltaic systems are summarized well by detailed facies models that relate processes of flow, sediment transport, and deposition. Facies models relating depositional processes to the facies distribution and architecture of tidally-influenced river deltas are less developed. Improved facies models of tidally-influenced river deltas require a clearer understanding of the interaction of river currents that supply deltaic sediments to the shoreline and the reworking of this sediment at the coast by tidal currents. Tidal influences on deltaic deposition can change the shoreline morphology, relationships between distributary channels and deposition under associated river plumes, the shape of prograding delta front sediment lobes, and depositional patterns on delta plains. Identifying these differences improves paleoenvionmental interpretations and predictions of subsurface variations in deltaic deposits that control aquifer and hydrocarbon reservoir heterogeneities.

Tide-influenced deltaic deposits are known to be architecturally complex, which makes it difficult to define longer-term facies trends and allostratigraphic surfaces (Reading, 1982; Wright, 1985; Willis, *in press*). Most comparisons between tide-

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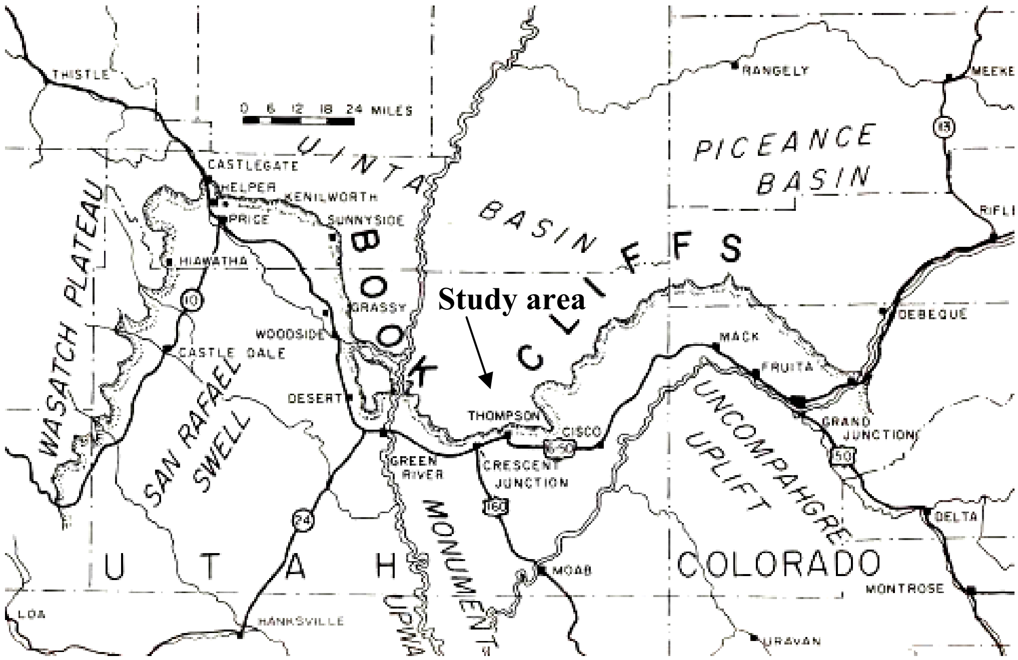



**Figure 1–Outcrop pattern of the Book Cliffs. This study of the upper Sego Sandstone extends between the towns of Crescent Junction and Thompson Springs, Utah (Young, 1966).**

dominated deltas and those of other types of deltas have focused on smaller-scale facies differences like the degree of heterogeneity, evidence for flow reversal, effects of tides within distributary channels and interdistributary delta top areas, and the elongation of sand bars formed on the delta front (Willis and Gabel, 2001). There are fewer studies that have addressed larger-scale differences in facies architecture and preservation formed as tide-influenced deltaic systems evolved during major transgressions and regressions of shorelines. Interpretation of longer-term controls on facies architecture requires that smaller-scale facies variations be mapped in detail over distances extensive enough to show regional changes in depositional patterns.

This study presents a survey of facies, stratigraphic trends, and continuous allostratigraphic surfaces across an 8.5 kilometer long outcrop of the upper Sego Sandstone, exposed in the Book Cliffs of eastern Utah (Figure 1). Changes in depositional processes recorded within the upper Sego Sandstone are interpreted to define the evolution of depositional environments. Criteria to distinguish larger-scale facies trends and stratigraphic surfaces from local depositional variations are identified to define a sequence stratigraphic framework. Previous studies of the lower Sego Sandstone have lead to contrasting paleoenvionmental and stratigraphic interpretations. Initial sequence stratigraphic interpretations of the Sego Sandstone suggested deposition within estuarine incised valley fill systems (e.g., Van Wagoner et al., 1991), following widely adapted assumptions that all tidal facies record transgressive sequence stratigraphic settings. More recent interpretations of the lower Sego Sandstone suggested these deposits formed dominantly on deltas that prograded into a shallow marine basin and had their tops ravined during transgression (Willis and Gabel, 2001). Although the upper Sego Sandstone is known to have many facies similarities with the lower (Van Wagoner et al., 1991; Willis, 2000; Yoshida et al., 2001), the facies architecture of this unit has not been studied in the same detail as the lower Sego Sandstone. Results of this study are compared with facies variations reported in similar studies of lower Sego Sandstone and modern delta deposits to determine if this unit can also be understood within the context of facies models for prograding tide-influenced deltaic shorelines (Van Wagoner et al., 1991; Willis and Gabel, 2001, Yoshida et al., 2001).

## DELTAIC FACIES MODELS

Delta morphology depends on the discharge and grain size of fluvial supplied sediment and the strength of basinal currents that can rework this sediment along the coast. Deltas can be divided into three major depositional environments: delta plain, delta front, and prodelta (Allen and Chambers, 1998). Deposits formed in each of these environments are characterized by different sand/shale ratios and sediment bodies with contrasting geometry and internal bedding (Van Wagoner et al., 1991).

Prodelta sediments, deposited basinward of the delta front, overlay marine inner shelf sediments. These laminated clays and silts rain out from suspension and are deposited at relatively slow rates. Slow rates of deposition allow extensive infaunal burrowing. The abundance and diversity of trace fossils decreases upward within these deposits as water shallows, sedimentation rates and depositional energy increases, and waters become more brackish towards areas of fluvial discharge (Wright, 1985).

The delta front includes the shoreline and steeply seaward dipping submarine portions of the delta (Reading, 1982). Facies variability increases and lateral continuity of sediment bodies decreases, relative to prodelta deposits. Delta front deposits coarsen, become better sorted, and less bioturbated upward as the delta progrades (Wright, 1985). Depending on the dominant coastal processes, distal bars may grade upward into distributary mouth bars, tide reworked ridges, or wave-reworked shoreface deposits, each containing beds with varying lithology and internal stratification (Wright, 1985).

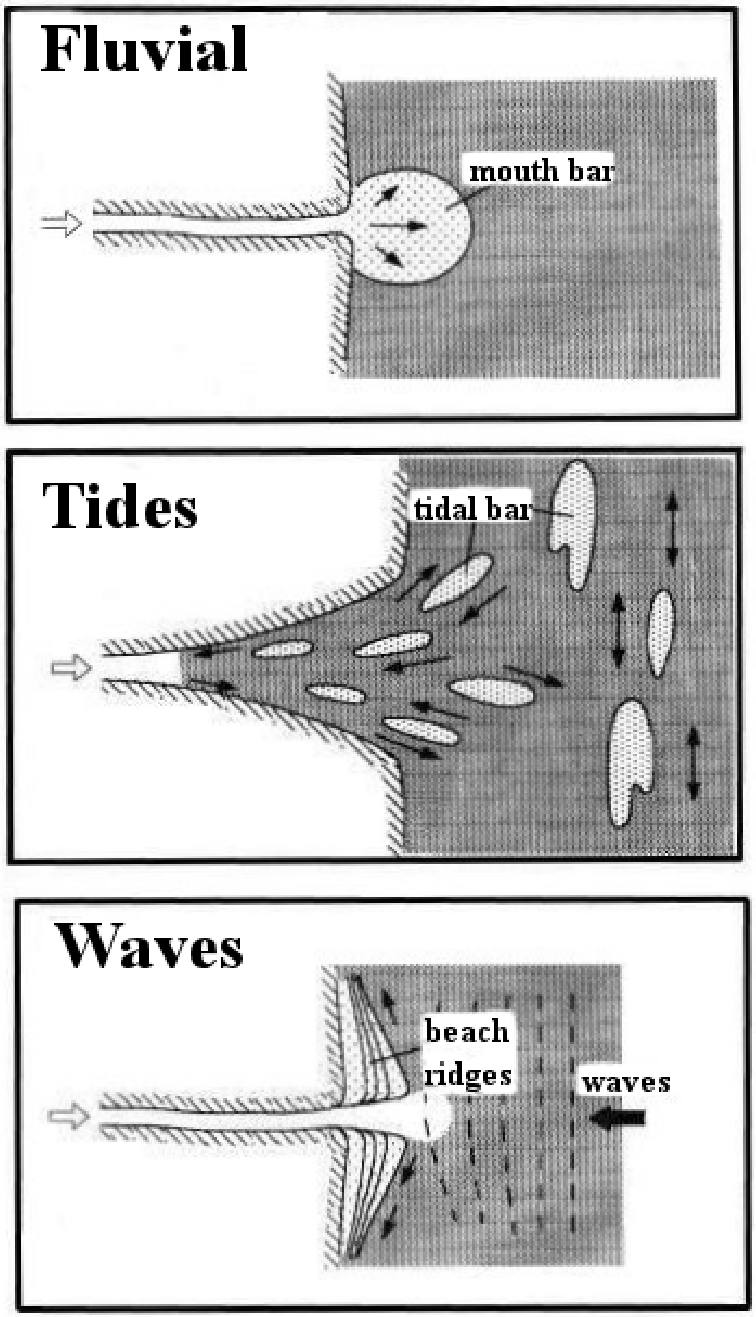



**Figure 2–Types of deltas based on the relative strength of different basinal and fluvial currents, which influence the shape of the coastline and sand bodies (Allen and Chambers, 1998).**

Delta plains, the subaerially exposed portion of the delta, are affected by fluvial and tidal processes, but seldom by waves (Reading, 1982). Distributary channels (extensions of the fluvial system) branch basinward as they divide fluvial discharge across the delta top to the delta front. Channel abandonment leads to the creation of an abandonment facies marker. Channel fill type and bioturbation depend on the amount of tidal influence and the discharge and grain size of the fluvial sediment supply. As sediment aggrades, distributary channel paths can diverge by as much as 60º from the average basin slope direction before they terminate at distributary mouths into bars at the delta front. Levees form adjacent to channels during river floods. When levees fail, crevasse splays are deposited within interdistributary areas. Interdistributary deposits are generally fine-grained accumulations formed in bays, floodplains, lakes, tidal flats, marshes, swamps, and salinas that are highly-bioturbated by fauna and vegetation (Reading, 1982; Wright, 1985).

Most deposition on river-dominated deltas occurs during major river floods, when unidirectional flows expand from distributary channels and deposit lobate mouth bar sand bodies onto the delta front (Reading, 1982; Allen and Chambers, 1998; Figure 2). Distributary channels avulse to regain gradient advantage as mouth bars grow, leaving clay-filled channels and abandoned mouth bars (Postma, 1995). Once abandoned, mouth bars tend not to be reworked, but rather become covered by muds deposited from suspension. On the delta top, sandy levees along distributary channels pass laterally into heavily bioturbated muds deposited within interdistributary bays (Reading, 1982). Banks of distributary channels locally overflow during river floods, depositing sandy crevasse splay sands, formed where flood water locally breach levees, farther into interdistributary areas (Figure 3). The prograding delta deposit is an upward-coarsening succession of stacked offset lobate mouth bar sands separated by thinner abandonment marine muds that pass upward into deposits of interdistributary bays and distributary channels (Reading, 1982).

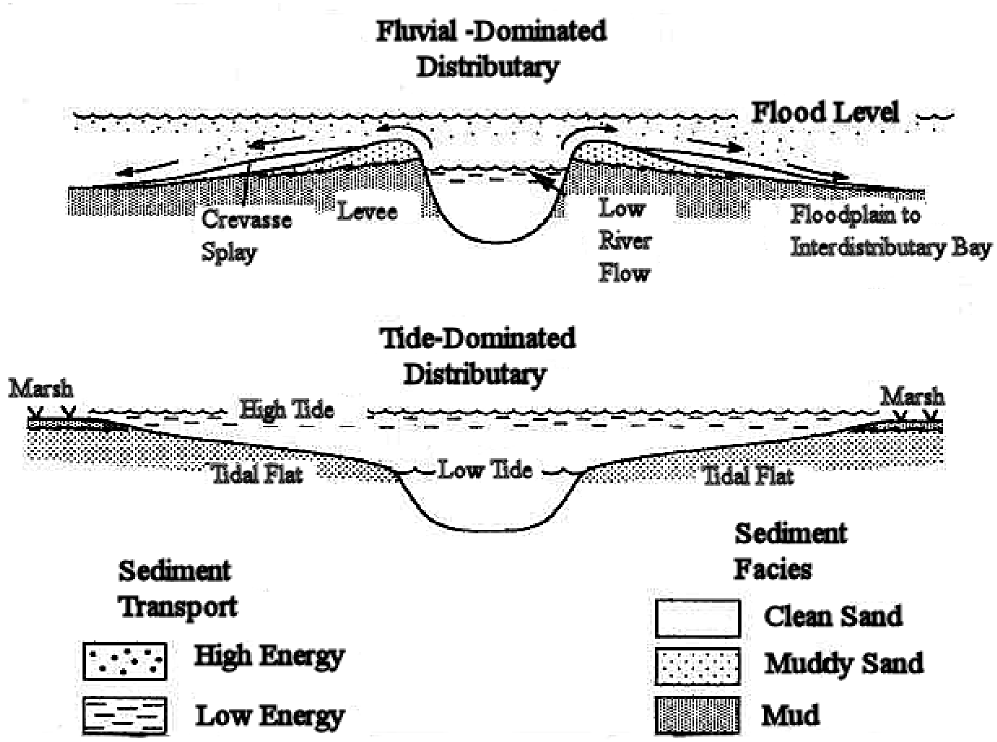



**Figure 3–Distributary channels differ in architecture depending on whether they are dominated by fluvial or tidal processes. Fluvial distributaries overflow during alluvial floods creating fine-grained overbank deposits and channel edge levees. Tidal distributary channel morphology is controlled by low and high tidal ranges (Allen and Chambers, 1998).**

On deltaic coastlines where tidal range is greater, tidal currents can intrude several tens of kilometers into distributary channels and affect the out-flowing distributary plume (Postma and Nemec, 1995). Diurnal tides rapidly change current speeds, resulting in more heterolithic deposits. Mouth bars deposited during river floods can be significantly reworked by tidal erosion and redeposition. Tide-dominated distributary channels have low sinuosity, high thickness/width ratios, and lack levees (Reading, 1982). Tidally-influenced channel deposits may fine upward as they are increasingly bioturbated (Reading, 1982). Flood tides enter distributary channels through funnel shaped mouths and then inundate interdistributary areas creating tidal flats. Flood tidal currents initially accelerate into interdistributary areas before gradually decelerating, reworking mud preferentially further landward. Linear tidal sand bars commonly occur in the lower reaches of distributary channels. These sand bars have been described along coastlines of the Gulf of Papua, the Mekong, the Mahakam, Ganges-Brahmaputra, and the Irawaddy river deltas. Although these bars have many similarities with distributary mouth bars in fluvial-dominated systems, contrasts in flow processes may leave recognizable differences in relationships between bed geometry, sedimentary structures and grain sorting. Fluvial mouth bars aggrade as flow is diverted away from bar crests and decelerates as it expands out of the channel. The result is a bar that coarsens upward and has internal beds that dip in the direction of expanding flows. Tidal bars in this setting may nucleate on mouth bars but continue to grow and become more elongate due to the acceleration of flows obliquely toward bar crests, which tends to increase sediment transport to the bar lee. Tidal bars can be hundreds of meters wide, several kilometers long, and 10-20 meters high. Bars are aligned nearly parallel to the channel and aggrade by lateral to oblique accretion (Reading, 1982; Willis, *in press*). Tidal sand ridges tend to migrate constantly in one direction, rather than switch direction due to minor fluctuations in the flow regime. The channel on one side of a bar may transport sediment

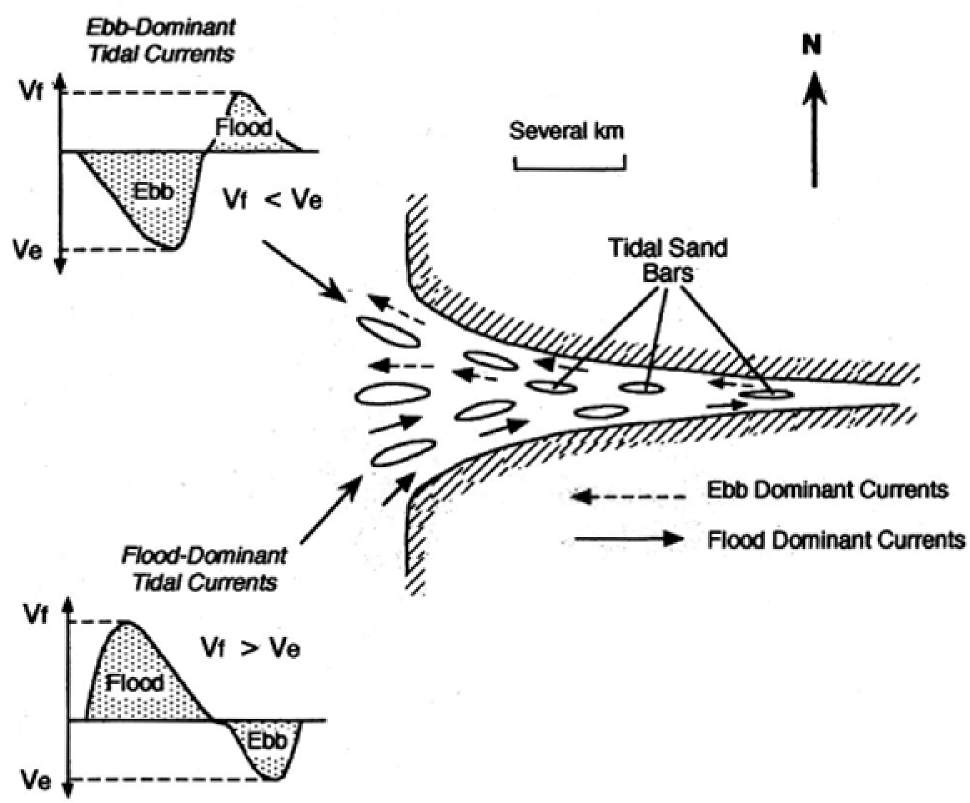



**Figure 4–Ebb and flood tidal currents vary in dominance across a tide-dominated delta. The resultant deposit will depend upon the dominant flow in the particular channels. Recent studies by Harris et al. (2004) shows that the dominant direction, ebb or flood, has little to do with the progradational nature of the delta deposit (Allen and Chambers, 1998).**

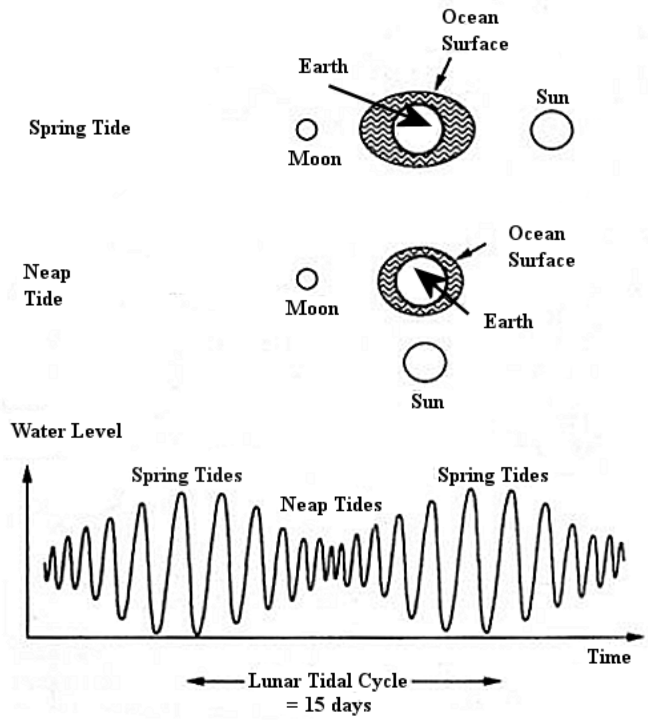



**Figure 5–Tides are largely the result of earth gravimetric interactions with the sun and the moon. A full lunar tidal cycle is usually 15 days. Within this cycle the tidal range varies on a daily basis gradually growing to the highest spring-tidal range. The lowest tidal range of the cycle is the neap tide. Neap tides occur when the sun and moon are perpendicular to the earth. Spring tides happen when the sun and the moon are aligned opposite each side of the earth (Allen and Chambers, 1998).**

preferentially in the ebb direction, while that on the other is dominated by flood currents with net sand transport in the opposite direction. Lobate mouth bars originally deposited by the river are thus extended by these diverging currents into linear sand bodies perpendicular to the shore (Reading, 1982; Willis, *in press*). As a result tidal sand bar deposits will show variations in current direction depending upon the direction that they locally grew under the prevailing currents. Deceleration of reversing tidal currents both offshore and onshore sustains the accumulation of sandy deposits along the coastline (Dyer et al., 1999; Figure 4). Bidirectional currents thus effectively hold sands within the distal mouths of tidal channels, where they can be reworked by diurnal tidal currents. A maze of channels and bars can extend across shallow delta top areas up to 95 km from shore, as in the case of the Ganges-Brahmaputra (Reading, 1982). Abandonment of an area of the delta top following distributary channel avulsion is likely to lead to longer periods of tidal reworking, because sediments are more slowly buried by fluvial sediment discharge (Willis and Gabel, 2003).

Allen and Chambers (1998) list four processes that set tidal deposits apart from their fluvial-dominated deltaic counterparts: (1) cyclicity, (2) bidirectional flows, (3) asymmetry between ebb and flood currents, and (4) absence of high energy flood event beds observed in fluvial-dominated settings. Cyclicity is the result of frequent changes in current speed and direction over several scales of tidal variation, from diurnal to seasonal, which causes alternating deposition of sand and mud (Figure 5). The thicknesses of these deposits will vary according to the amount of sediment and the strength of the tidal regime during these cycles (Nio and Yang, 1991; Allen and Chambers, 1998). Double mud drapes, formed by dual high and low tide slack-waters, are a sedimentary structure distinctive of tidal deposits (Figure 6). Herringbone cross-stratification, recording the migration of bedforms in opposite directions under ebb and flood currents, separated by a period of still water, is less common because opposing tidal flows tend to move along different paths. Oppositely oriented ripples, superimposed on larger scale cross strata, are more common because dunes seldom fully form during single tidal cycles (Dalrymple and Rhodes, 1995). It is also common to observe reactivation surfaces within cross sets, which record the beveling back of dune fronts during reversed flows and then regrowth in the dominant flow direction (sensu Nio and Yang, 1991).

Tidal asymmetry in deposits occurs because of the differing power between the ebb and flood tides. At individual locations within tidally influenced areas, currents tend to be either ebb or flood dominant, but because of changing flow this dominance can change and superimposed flood or ebb deposits can form. Within river mouths the ebb currents would be expected to dominate because of the additive force of fluvial discharge. Rapid influxes of sediment, producing distinct depositional beds, are likely to reflect river floods rather than tidal reworking (Allen and Chambers, 1998).

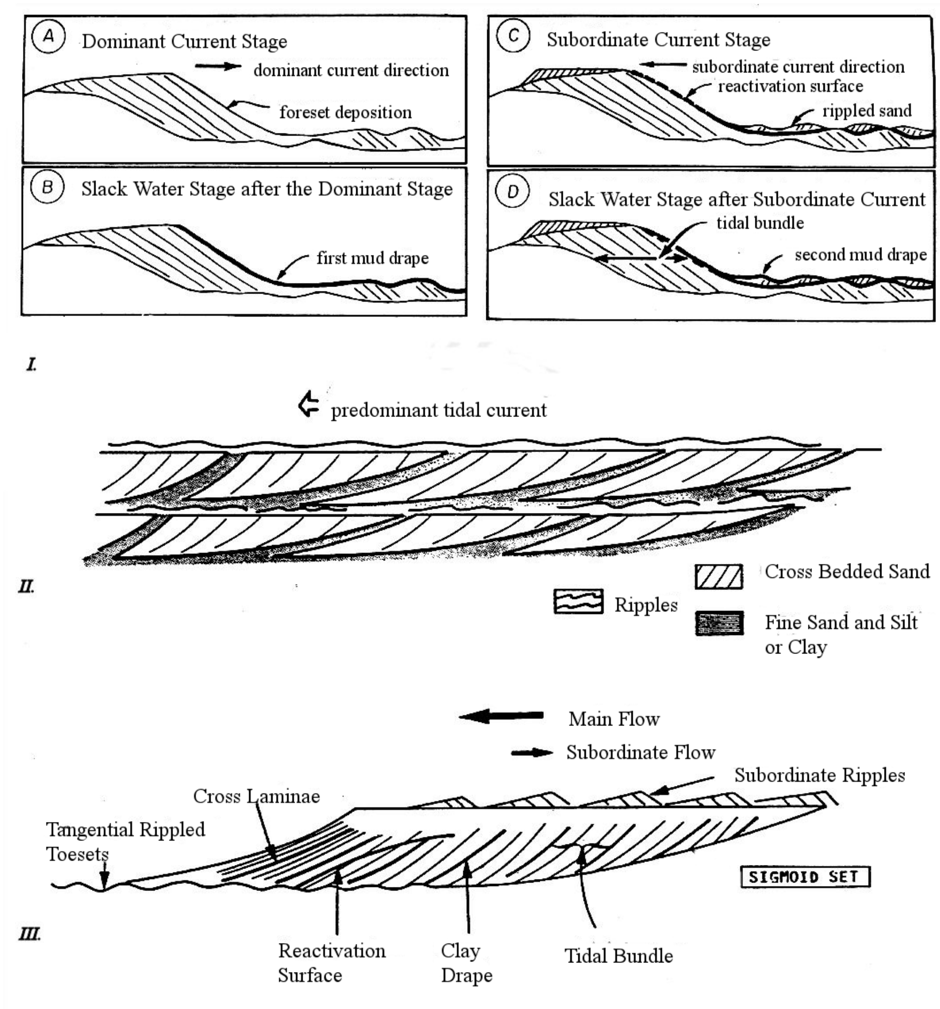



**Figure 6–Common sedimentary structures found in tidally-influenced environments. Some of these include: I) the formation of double mud drapes formed during reversals in flow direction; II) Sigmoidal sets signify asymmetrical tidal currents at several ranges of flow; III) A single sigmoid is composed of several tides at different scales of flow, internal drapes and reactivation surfaces show the influence of daily tides while the contacts between the larger sets show tidal variations on most likely a spring/neap tidal range (Allen and Chambers, 1998).**

# REGIONAL SETTING

The upper Sego Sandstone was deposited along the western margin of the Cretaceous Western Interior Seaway (Figure 7). This epicontinental sea connected the northern Boreal Sea with the Gulf of Mexico (Krystinik et al., 1995). The seaway formed in

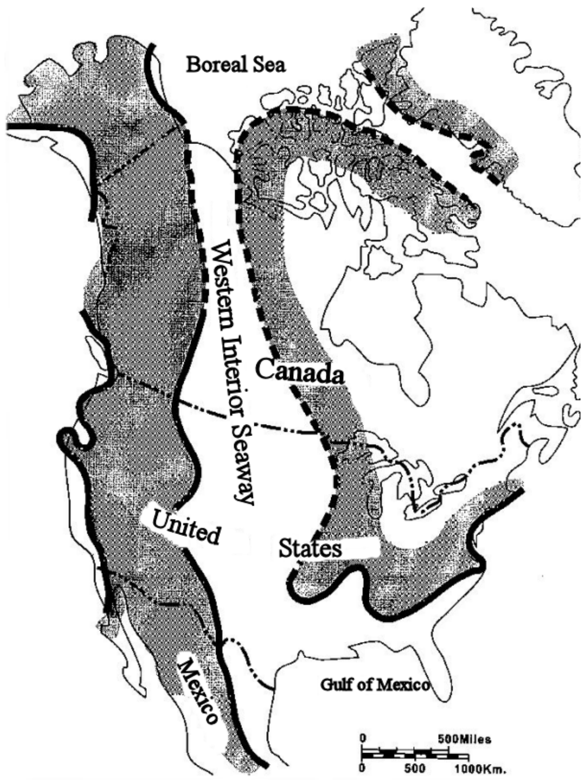



**Figure 7-The Western Interior Seaway flowed from Jurassic to Eocene time. In different stages of growth and decline, it connected the Boreal Sea with the Gulf of Mexico. This sea provided the basin and hydrodynamics necessary for the upper Sego Sandstone to form (Krystinik et al., 1995).**

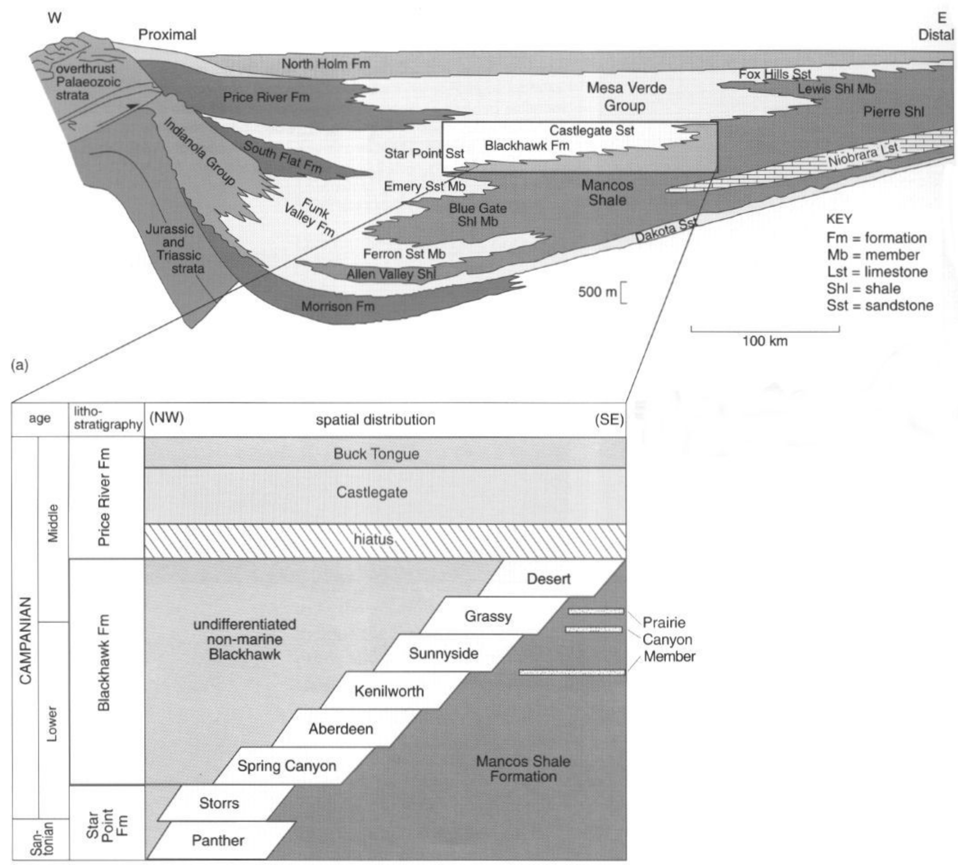



**Figure 8-A portion of the foreland basin strata deposited within the Book Cliffs. This diagram provides a reference for formational names and stacking patterns described in the text. Note the interfingering pattern of fully marine deposits laid down from the east with overall forward-stepping coastal and near-marine strata (modified from Coe, 2003).**

response to subsidence of a foreland basin parallel to the Sevier Orogeny (Figure 8). The upper Sego Sandstone was deposited at a relatively constant paleolatitude of 42º (Coe, 2003). The age of the Sego Sandstone is estimated to be Late Campanian based on ammonite age-dates (Gill and Hail, 1975). The ammonite *Baculites perplexus* found within shales of the upper Buck Tongue (Figure 9) directly under the Sego Sandstone is estimated to be ~76 mybp in age, and *Baculites scotti* found in shales of Anchor Mine Tongue within the Sego Sandstone is estimated to be ~ 74.6 mybp in age. This gives an age range of about 1.4 million years for the lower part of the Sego Sandstone (Gill and Hail, 1975; Van Wagoner et al., 1991).

The Sego Sandstone in central Utah, a member of the Mancos Shale, is one of numerous sandstone tongues that record the episodic progradation of shorelines eastward toward the axis of the Western Interior Seaway. Older sandstone tongues, which successively extend farther into the basin, are grouped into the Blackhawk and Castlegate Members of the Mancos Shale in central Utah. The Sego Sandstone is separated from these earlier sandstones by the Buck Tongue of the Mancos Shale, which records a major transgression of the seaway onto the underlying Castlegate sandstone. Although the upper part of the Buck Tongue contains progressively more thin beds of sandstone, the base of the Sego Sandstone is defined by a distinct erosion surface and a pronounced coarsening. The Sego Sandstone is divided into lower and upper layers by a second transgressive marine shale, the Anchor Mine Tongue of the Mancos Shale. Like the lower Sego Sandstone, the upper is defined by an abrupt coarsening above an erosion surface.

The Sego Sandstone passes upward into coastal plain deposits of the Neslen Formation, which aggraded behind a succession of vertically stacked, wave-dominated shoreface sandstones located along the Utah-Colorado boarder. The contact of the Sego with the Neslen Formation is defined by the transition from cliff-forming sandstones to slope forming muddier deposits with coals (Hettinger and Kirshbaum, 2002). Although this division is fairly distinct at a regional scale, it is less useful for detailed mapping because upper parts of the Sego Sandstone interval are locally very fine grained and basal parts of the Neslen Formation can contain thick sandy channel deposits. Higher within the Neslen, deposits become dominated by fluvial channel deposits that bypassed sediment to shorelines that continued to prograde into Colorado (the Blue Castle Tongue).

There are relatively arbitrary changes in stratigraphic nomenclature within the interval that includes the Sego Sandstone at two locations along the Book Cliffs: 1) At Green River, Utah, a few tens of kilometers west of this study area); and 2) at the Utah-Colorado State boarder a bit over 100 kilometers east of this study area (Franczyk, 1989; Hettinger et al., 2002; Figures 9). The Buck and Anchor Mine Tongues of the Mancos Shale both pinch out west of Green River, Utah, as the strata become sandier in more proximal areas of the basin. Even further west, in the area of Price, Utah, all the strata from the base of the Castlegate Sandstone through to the Blue Castle Tongue are defined to be members of the Castlegate Sandstone. Although a finer grained interval referred to as the middle Castlegate Member near Price, Utah, correlates in a general way to the Buck Tongue and Sego Sandstone members of the Mancos Shale in central Utah, details of this correlation are currently debated. Yoshida et al. (2001) suggest that this finer-grained middle Castlegate Member correlates within basal parts of the Neslen in central Utah, and suggest that the Sego Member is part of a larger-scale transgressive succession.

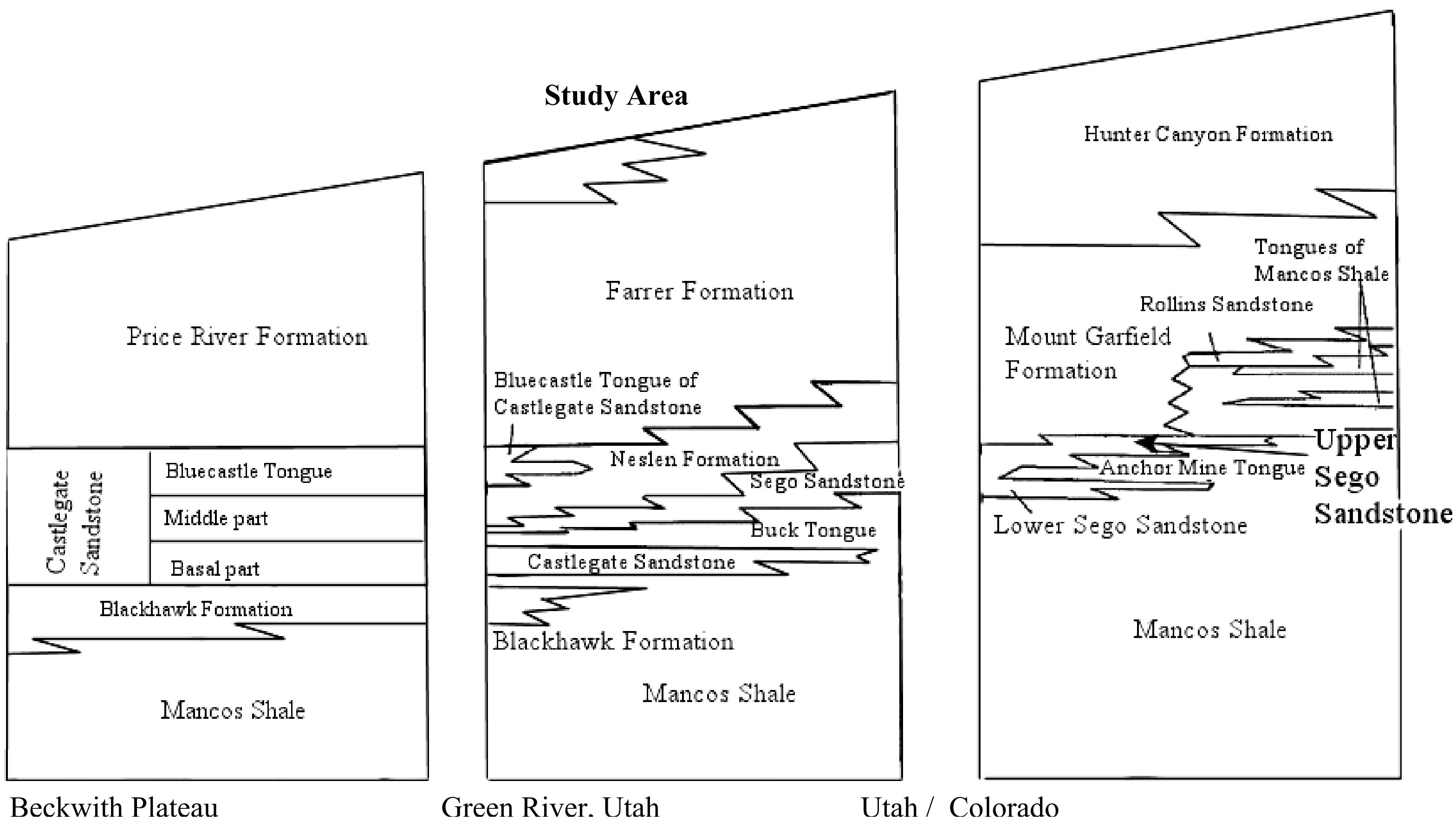



**Figure 9–Stratigraphic nomenclature of the Book Cliffs changes with geographic area, state boundaries, and facies markers. The Sego Sandstone changes in status down dip from its beginnings in the middle part of the Castlegate Sandstone. Before the Sego Sandstone gained formational status it was listed as part of the Price River Formation (Franczyk, 1989).**

McLaurin and Steel (2001), in contrast, suggest that the finer-grained middle Castlegate Member correlates with the Buck Tongue and thus they imply that the Sego Member in central Utah formed during renewed long-term regression. Although the Sego Sandstone of Utah pinches out to shale near the Colorado border, a different sandstone wedge in Colorado (the first one there above the Buck Tongue Shale) is also called the Sego Sandstone. The Neslen Formation of Utah correlates with the basal part of the Mount Garfield Formation in Colorado.

The Sego Sandstone is part of a finer-grained interval within the overall fill of the Cretaceous Seaway, sandwiched between clastic wedges that rapidly prograded into the seaway. Gradual westward (landward) thinning of the Buck and Anchor Mine Tongues formed by flooding across the top of the underlying Blackhawk-Castlegate clastic wedge and was likely a broad, shallow margin of the seaway. In such a setting, higher frequency transgressions and regressions may have been relatively extensive and tidal currents may have been enhanced by resonance. Interpretations of depositional environments of the Sego Sandstone changed over time. Franczyk (1989) interpreted the Sego Sandstone to be stacked shoreface deposits on a microtidal barrier island coastline, and related deposition to a relative rise in sea level caused by changes in basin subsidence to sediment supply. Van Wagoner et al. (1991) presented the first detailed sequence stratigraphic interpretations of the Sego Sandstone, and suggested the deposits contained nine high-frequency sequences. They suggested that regressive parts of each sequence were eroded during transgression, and that only deposits of estuarine transgressive valley fills were preserved. Willis and Gabel (2001; 2003) reexamined parts of the lower Sego Sandstone and suggested instead that these are deposits of tide-dominated river deltas that prograded onto the shallow margin of the seaway. Although they recognized the high-relief erosional surfaces that Van Wagoner et al. (1991) used to define valley fill sequences, they suggested that the lower Sego is composed of two dominantly regressive successions that were dissected by tidally enlarged distributary channels and perhaps very locally by estuarine-filled valleys. Kirschbaum and Hettinger (2004) recently presented evidence the distal-most end of the upper Sego Sandstone comprises a succession of falling stage tide-influenced shoreline deposits that underlie a vertical stack of wave-dominated shorefaces deposited within the Mount Garfield Formation.

## METHODS

The upper Sego Sandstone is exposed in nearly continuous vertical cliffs over the 9.7 km distance between Sego and Crescent Canyons in east central Utah. These strata dip northward at 2-4 degrees, and are not deformed, other than being cut by a few minor normal faults. This outcrop belt extends along a northeast-southwest trend, nearly perpendicular to the average southeast dispersal of Upper Sego sediments inferred from regional studies (Van Wagoner, 1998; Kirschbaum and Hettinger, 2004). Side canyons allow more limited documentation of variations along depositional dip. Thirty-two

**Table 1: GPS coordinates of sedimentary logs (measured section).**

| # | log ID | coordinates | # | log ID | coordinates |
|---|---|---|---|---|---|
| 1 | B2 | 12S 609794 4318500 | 17 | B6 | 12S 608485 4319273 |
| 2 | B4 | 12S 609342 4318868 | 18 | C1 | 12S 605518 4318997 |
| 3 | B5 | 12S 609012 4318868 | 19 | C2 | 12S 606249 4319132 |
| 4 | B7 | 12S 608170 4319121 | 20 | S1 | 12S 612157 4321341 |
| 5 | B8 | 12S 608007 4318866 | 21 | S2 | 12S 611683 4320533 |
| 6 | B10 | 12S 608045 4317872 | 22 | T8 | 12S 610568 4319910 |
| 7 | B11 | 12S 607780 4317803 | 23 | T9 | 12S 610701 4320070 |
| 8 | B12 | 12S 607304 4317936 | 24 | T1 | 12S 610888 4321090 |
| 9 | B13 | 12S 607160 4317634 | 25 | T2 | 12S 610520 4319853 |
| 10 | B14 | 12S 607034 4317310 | 26 | T3 | 12S 610734 4320061 |
| 11 | B15 | 12S 606799 4317082 | 27 | T4 | 12S 610768 4319136 |
| 12 | B16 | 12S 606505 4316954 | 28 | T5 | 12S 610557 4318919 |
| 13 | B17 | 12S 606417 4316396 | 29 | T6 | 12S 610558 4321339 |

**Table 1: Continued.**

| # | log ID | coordinates | # | log ID | coordinates |
|---|---|---|---|---|---|
| 14 | B18 | 12S 605466 4317358 | 30 | T7 | 12S 610814 4319462 |
| 15 | B19 | 12S 605374 4316653 | 31 | T10 | 12S 610310 4319090 |
| 16 | B20 | 12S 605644 4316235 | 32 | T11 | 12S 610118 4318968 |

vertical section was measured with Jacob staff, Abney level, and meter tape record grain size, sedimentary structures, and ichnofossils (Table 1; Figure 10). The degree of bioturbation was estimated on a scale of 1 through 5, with 1 indicating very-sparse burrows and 5 indicating complete obliteration of depositional stratification. Positions of logs were recorded on topographic maps and using GPS. Technical climbing gear and ropes were used to access areas exposed in vertical cliffs. Logs, spaced on average about 1/3 km apart, extend from the Anchor Mine Tongue to the first thin coal bed within the overlying Neslen Formation. Lithic variations defining bedsets and key stratal surfaces (including distinctive bedsets, erosion surfaces, continuous thin shales, and cemented beds) were traced in the field between logs across continuous outcrop exposures. Photomosaics of selected outcrops aided correlation. Measured log positions and correlated surfaces were projected normal to a best-fit line (oriented 53° east of north) to construct a bedding diagram that shows the architecture of facies changes between different types of stratal surfaces (Figure 11 and Figure 12).

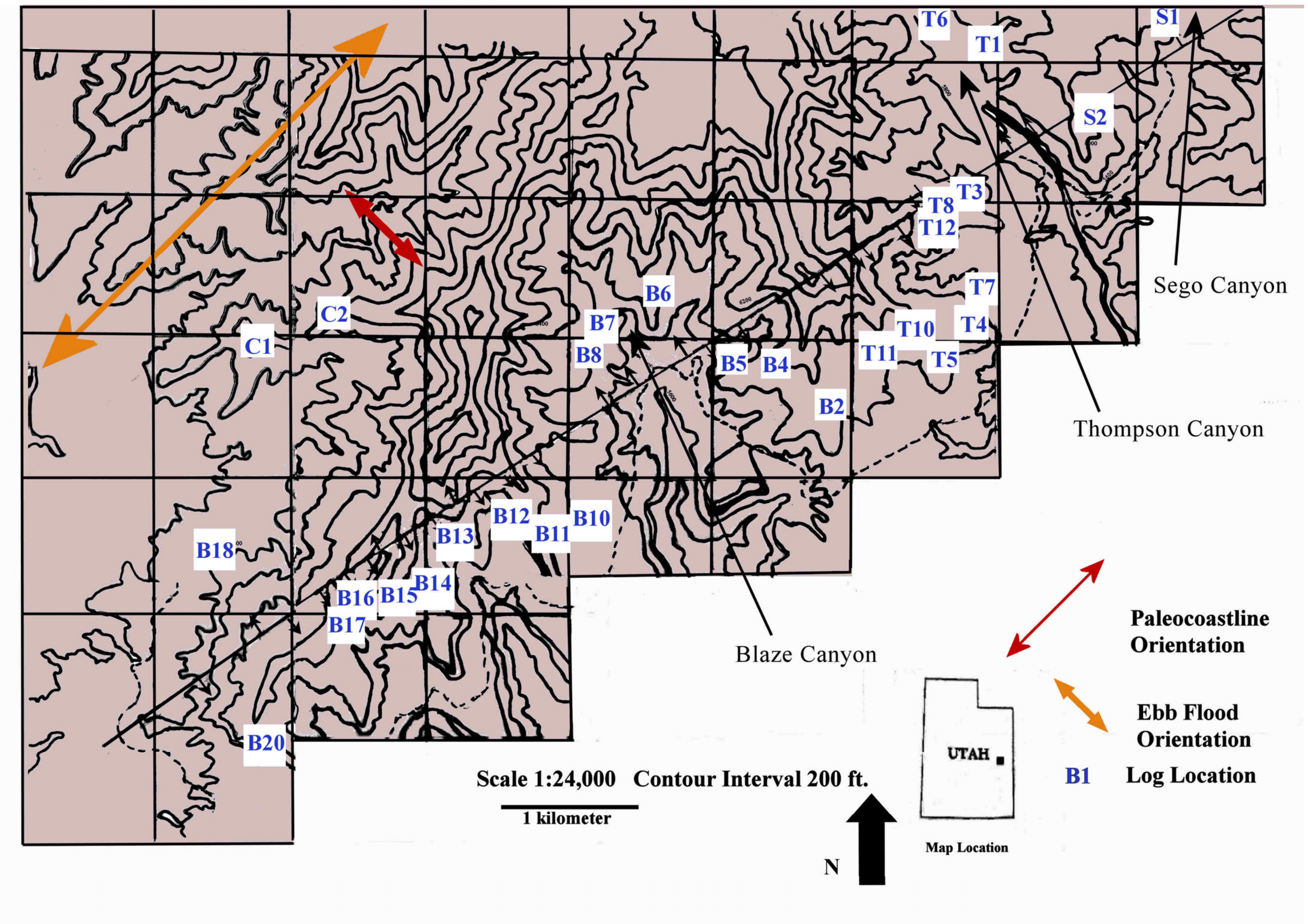



**Figure 10–Map shows locations of sedimentary logs measured during this survey. Paleocoastline and inferred ebb-flood orientations are included for reference (modified from United States Geological Survey, 1991a; United States Geological Survey, 1991b; United States Geological Survey, 1991c).**

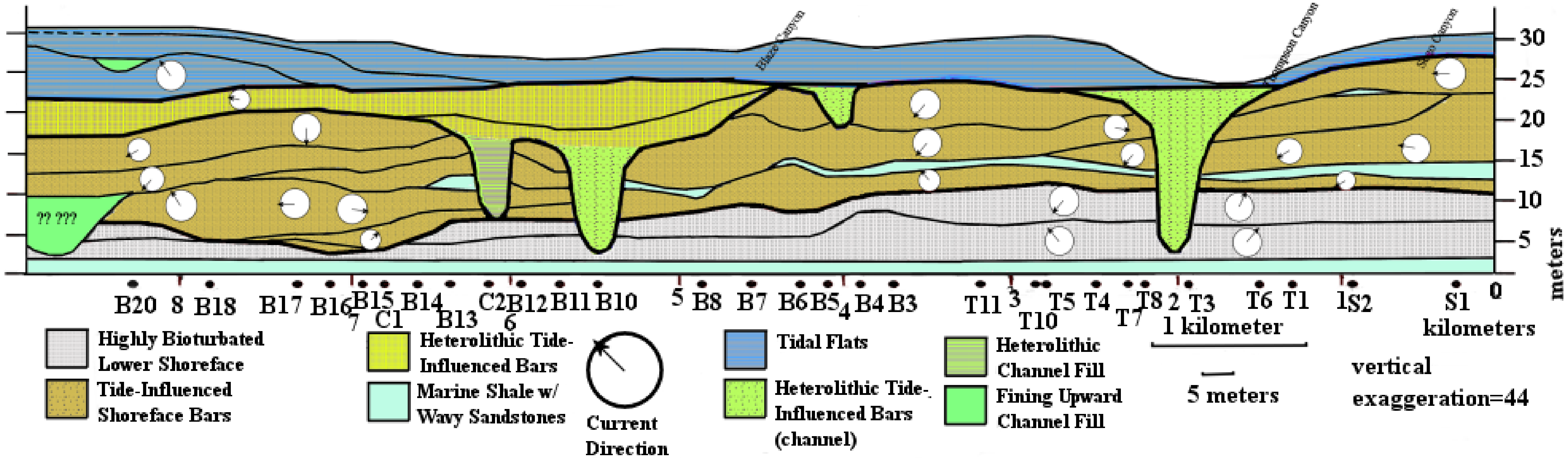



**Figure 11-The final bedding diagram for the Campanian upper Sego Sandstone displays the stacking pattern of the sediments as defined by abrupt facies transitions and erosion surfaces. The cross-section defines an 8.5 kilometer section that begins in Sego Canyon outside of Thompson Springs,**

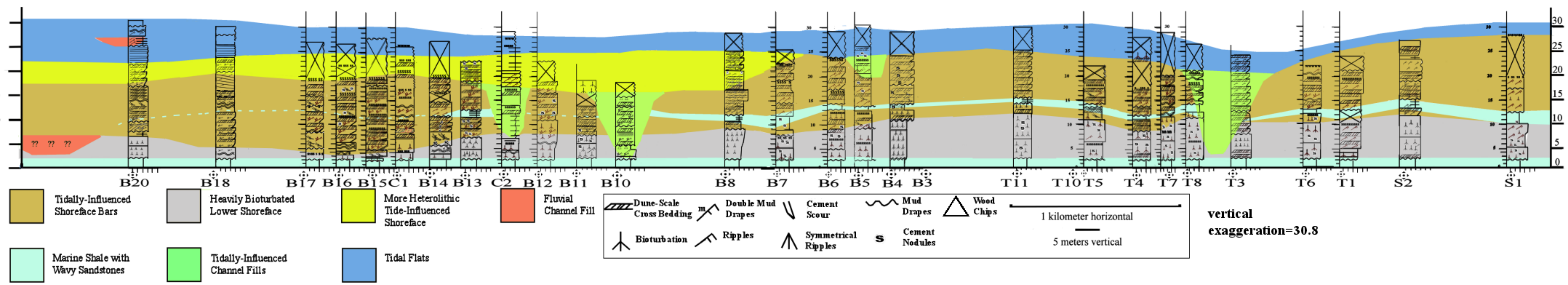



**Figure 12-Bedding diagram of major facies transitions with sedimentary logs overlain.**

# SEDIMENTOLOGY

Upper Sego Sandstone deposits can be separated into five different facies associations: 1) marine shale with wavy sandstones, 2) highly-bioturbated lower shoreface, 3) tidally-influenced shoreface bars, 4) tidal flat deposits, and 5) heterolithic tidal deposits. Three different subdivisions of diagenetic overprinting is also recognized. These facies and their internal geometry are described and interpreted first. In later sections, the facies geometry and stratigraphic variations are presented.

## Marine Shale with Wavy Sandstones

### Description

Beds of these dark grey to black, silt- and clay-sized mudstones range form less than one meter to several meters thick. Thin interbedded sandstone beds locally contain

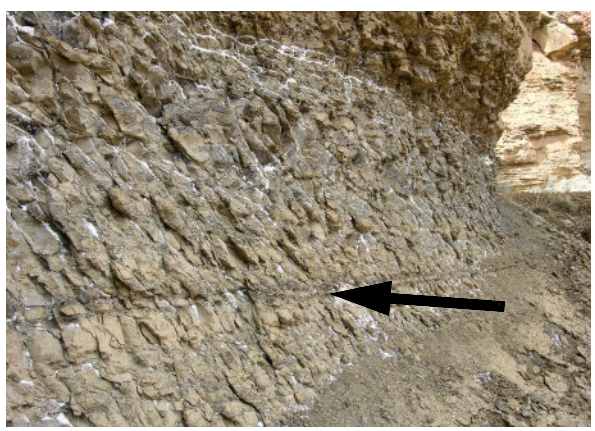


**Figure 13-Anchor Mine Tongue marine shale is usually exposed in an undercut bed below the upper Sego Sandstone. Grey shales contain hummocky stratified sandstone (arrow).**

symmetrical ripples and small sets of hummocky cross stratification (Figure 13). Sandstone beds are generally continuous over several hundreds of meters to kilometers, but can abruptly pinch out along strike. Beds increase in sand upward but abruptly change to shale at their tops. Bioturbation is extensive. Cemented nodules, shale rip-up clasts, and wood fragments are common, particularly near the base of beds.

Interpretation

These deposits typically form in quiescent offshore zones where clays and organic material settle from the water column below storm wave base (Kirschbaum and Hettinger, 2004). Intense bioturbation usually reflects open-marine conditions (Pemberton et al., 2001). Black shales form due to sediment starvation and concentration of organic clays (Willis and Gabel, 2001), whereas grey shales are more oxidized muds churned by the infauna. Although some of these deposits may form from the deposition of fluid muds moving down slope away from distributary plumes (Harris et al., 2004), this is difficult to demonstrate because of subsequent bioturbation. Shale rip-up clasts, wood, and siderite nodules indicate sea-floor scouring during storm events that passed sands basinward (Willis and Gabel, 2001). Thin Sandstone beds form by wave reworking of the sea bed during major storms (Boggs, 1995; Kirschbaum and Hettinger, 2004).

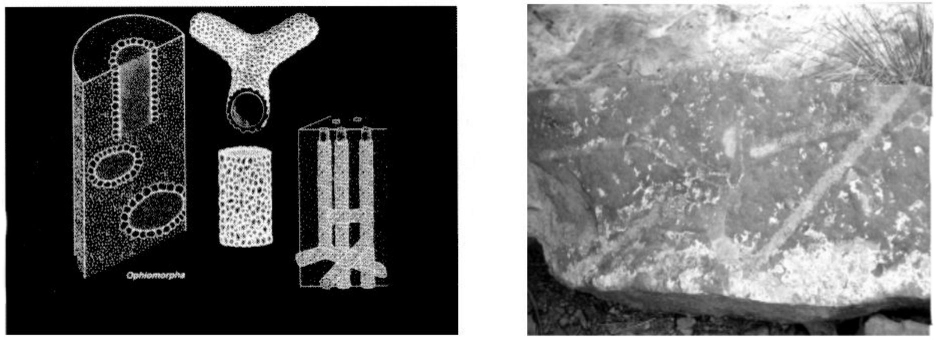
**Highly-Bioturbated Lower Shoreface**

Description

These lower-very-fine- to fine-grained sandstone beds vary in thickness from 1 to 6 meters and are continuous over kilometers (Figure 14). Although sedimentary structures are generally obscured by intense bioturbation (level 5) and overprinting of cements, poorly preserved small-scale cross beds are observed. Symmetrical ripples and hummocky cross-stratification are also locally observed near the base of beds locally. Internal, discontinuous, lenses of mud are commonly replaced with cement. Rare soft sediment deformation features and decimeter-relief erosion surfaces are commonly


**Figure 14-Heavily bioturbated sandstones that overlie the Anchor Mine Tongue make up the lower to middle shoreface deposits. Extensive bioturbation of sediments is linked to slow deposition rates.**

overlain by cement replaced mud. Trace fossils are dominated by *Ophiomorpha*, but also include *Chondrites*, *Paleophycus*, *Rosselia*, and *Planolites* (Figure 15; Figure 16). Discontinuous layers and nodules of cement are common.



*Ophiomorpha*

**Figure 15–*Ophiomorpha* is a ubiquitous trace fossil found within nearly all facies of the upper Sego Sandstone. The trace is thought to be result of the burrowing Thalassinoide shrimp known to infiltrate all near-marine environments. The main diagnostic feature of the trace burrow is the pelleted wall (modified from Pemberton et al., 2001).**

Interpretation

These sheet-sandstones formed in middle to lower shoreface zones (Figure 17). Rare hummocky cross stratification records storm activity below fair-weather wave base, and ripple cross stratification suggests at least local unidirectional current flow, perhaps associated with storm setup return currents of tides. The complete mixing of the sediment by *Ophiomorpha* suggests slow depositional rates. The size and diversity of trace fossils suggests fully marine waters. Discontinuous cements scattered throughout are characteristic of reworked marine sands (Taylor, 2000).

**Tidally-Influenced Shoreface Bars**

Description

These sandy bedsets vary from a meter to over ten meters thick, and extend laterally for hundreds of meters to several kilometers. Bedsets generally have an erosional base,

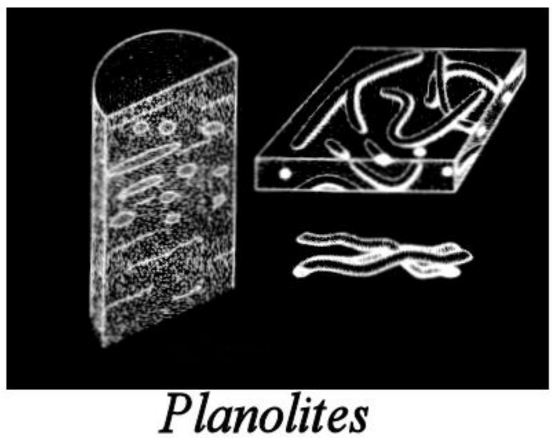


**Figure 16–*Planolites* is possibly the trace of Sipunculid. These complex root like burrows are part of the feeding strategy of an organism typically associated with the Cruziana ichnofacies Pemberton et al., 2001).**

internal inclined beds, and coarsen upward. Individual beds thicken upward within bedsets. Beds dip more steeply higher within sets (up to 5-10 degrees) and tangentially decrease in dip downward into the basal erosion surface. Bed dips are steeper on average where bedsets are thicker. Although average bed dip directions vary laterally along bedsets, and between different examples, bed dips within most sets average northwest and within a few sets average southeast. The tops of bedsets are often cemented,

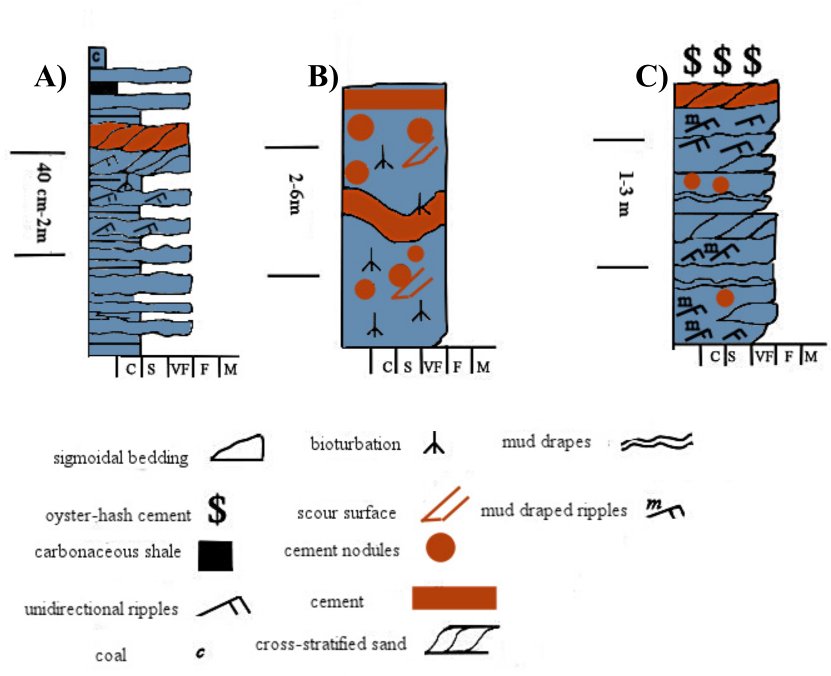



**Figure 17–The three main facies groups represented in the upper Sego Sandstone are: A) tidal flat deposits/intertidal zone, B) highly bioturbated middle/lower shoreface bars, and C) tide-influenced shoreface bars. Each sandstone has a distinctive facies that represent differing water depths and flow regimes.**

extensively burrowed by *Ophiomorpha* (Figure 18), and rarely defined by distinctive lag deposits of oyster shells.

Basal deposits within bedsets contain centimeter-thick, very-fine- to fine-grained sandstone beds, thicker mud drapes, and abundant shale rip-up, wood, siderite-nodule clasts. Small-scale cross stratification dominates, and it can dip in opposite directions. Large-scale cross strata within some beds can be sigmoidal, contain abundant reactivation surfaces, and locally contain internal small-scale cross strata that climb up the larger-scale strata. High-angle, discontinuous scours at the base of some cross sets are cemented. Thick mud drapes can be massive or bioturbated. Bioturbation can increase upward within beds and varies overall from sparse to levels of 2-3 where disrupted by *Rosselia*, *Chondrites*, *Paleophycus*, and *Planolites*. Sandstone beds are thicker, amalgamated and dominated by nearly angle-of-repose, large-scale cross stratification

**Figure 18–This sandstone underlies the flooding surface of the first transgressive sequence. Note the abundant *Ophiomorpha* on the surface.**

higher within sets. Planar strata near the base of beds are rare. Internal mudstone drapes are sparse and discontinuous. In some cases cross strata in these thicker beds dip southeast, even in beds dominated by northwest dipping cross strata overall.


**Figure 19–Tidal bedsets record the progradational nature of the delta. These bedsets consist of small scale beds recording the daily interactions of tides grading upwards into thicker sands recording the progradation of distributary mouth bars.**

Interpretation

Bedsets form by the progradation and abandonment of tidal distributary mouth bars in the subtidal area. The dimensions of these bedsets are similar to the dimensions of tidal bars on modern tide-influenced deltas and estuaries (Off, 1963; Dalrymple, 1984; Coleman et al., 1988; Allen and Chambers, 1998; Li et al., 2002). Inclined beds, which formed during periods of high-velocity flow and then quiescence, also show the bar geometry in the direction of migration. Beds within sets that dip toward the northwest record dominantly flood accretion of a bar, whereas those toward the southeast accreted seaward.

Shale rip-up clasts and wood fragments along the basal surfaces accumulated when faster currents eroded the delta front. Upward coarsening and thickening of beds within sets records progradation of the distributary mouth bar in the tidal environment (Kirschbaum and Hettinger, 2004; Figure 19). Heterolithic deposits in basal parts of bedsets reflect high suspended sediment concentration. Abundant mud drapes, reactivation surfaces within cross sets, and oppositely dipping cross strata within adjacent sets are evidence for tidal influence on deposition. Patterns of mud drapes record the daily interaction of tides, and in the longer term, the cyclicity of spring and neap tides (Nio and Yang, 1991; Figure 20). Recognition of fluid mud in outcrop is difficult

**Figure 20–This photo from the study area shows a common tidal feature to the upper Sego Sandstone tidal bedset facies. The double mud drape evolved during times of reversing flow. During slack-water periods between the two dominant tides a mud drapes settles due to waning flow.**

because both fluid muds and mud settling from suspension can be quickly reworked by organisms (Wheatcroft and Drake, 2003) Increasingly high-angled cross stratification, fewer shale drapes, and coarse grain size records faster flow rates and increased sediment supply across proximal parts of bars. Rare planar stratification with parting lineation at the base of some beds was deposited during peak flow periods. Increased bioturbation at the top of a deposit may reflect a period of quiescence where little sediment was deposited allowing infaunalization. Amalgamation of some bedsets may reflect erosion during transgressive reworking of sands by tidal currents preceding progradation a subsequent bar.

Estimates of paleobathymetry are difficult because tides can build similar sandy bedforms in a wide range of water depths. The thicknesses of bedsets are inferred to decrease with local rates of sediment supply. Thus thickest and least heterolithic bedsets are interpreted to form near distributary mouths and thinner more-bioturbated bedsets to form further from areas of direct sediment input (Reading, 1982; Harris et al., 2004). Although deposition rates were probably also higher in more proximal deposits, this is difficult to document because of short-term variations expected in this environment. Low bioturbation in the majority of this facies indicates deposition was fairly rapid. Changes in the rate of riverine sediment discharge may be an important control on bar preservation. In the Han River Delta, up to 2 meters of sand can be deposited during individual river floods, 5 centimeter beds can be deposited in a single tidal cycle, and 40 centimeters of sediment accumulated during individual neap-spring cycles (Choi et al., 2004). Long-term preservation of tidal bundles is rare so extracting a depositional time frame is difficult. Accumulation of tidal sediment is not constant over long intervals due to the erosional capabilities of wave and wind-driven currents (Choi et al., 2004).

Sedimentary structures record principally flood-oriented current directions, even in those bedsets with ebb-dipping beds. It is a common observation that current dominance varies on opposite sides of tidal bars, reflecting varying directions of tidal flow onshore and offshore (Harris, 1989; Dalrymple and Rhodes, 1995). Harris (1989) stated that there were two possible reasons for tidal asymmetry: a change in the major current direction or the additive properties of wind-driven currents. A number of studies have discussed preferential preservation on flood oriented bedforms in larger scale prograding systems or dune reorientation over short time-scales (Berne' et al., 1993; Harris et al., 2004; Choi et al., 2004).

**Heterolithic Tide-Influenced Shoreface Bars**

Description

Bedsets are broadly similar to those in the Tide-Influenced Shoreface Bar facies, but are only a few meters thick and tend to be muddier. Unfortunately this facies tends to be poorly exposed because cliffs of muddy deposits tend to slump and become deeply weathered. In some cases, black fissile shale grades upward into heterolithic interlaminated shales and wavy sandstones with reversing small-scale cross stratification. In other cases bedsets grade upward from wavy sandstones and mudstones to amalgamated decimeter-thick flaser-bedded sandstones. Some bedsets fill small erosional scours, decimeters deep. Where bioturbated, cements are pervasive. Although cements make it difficult to identify trace fossils, *Ophiomorpha* is clearly common.

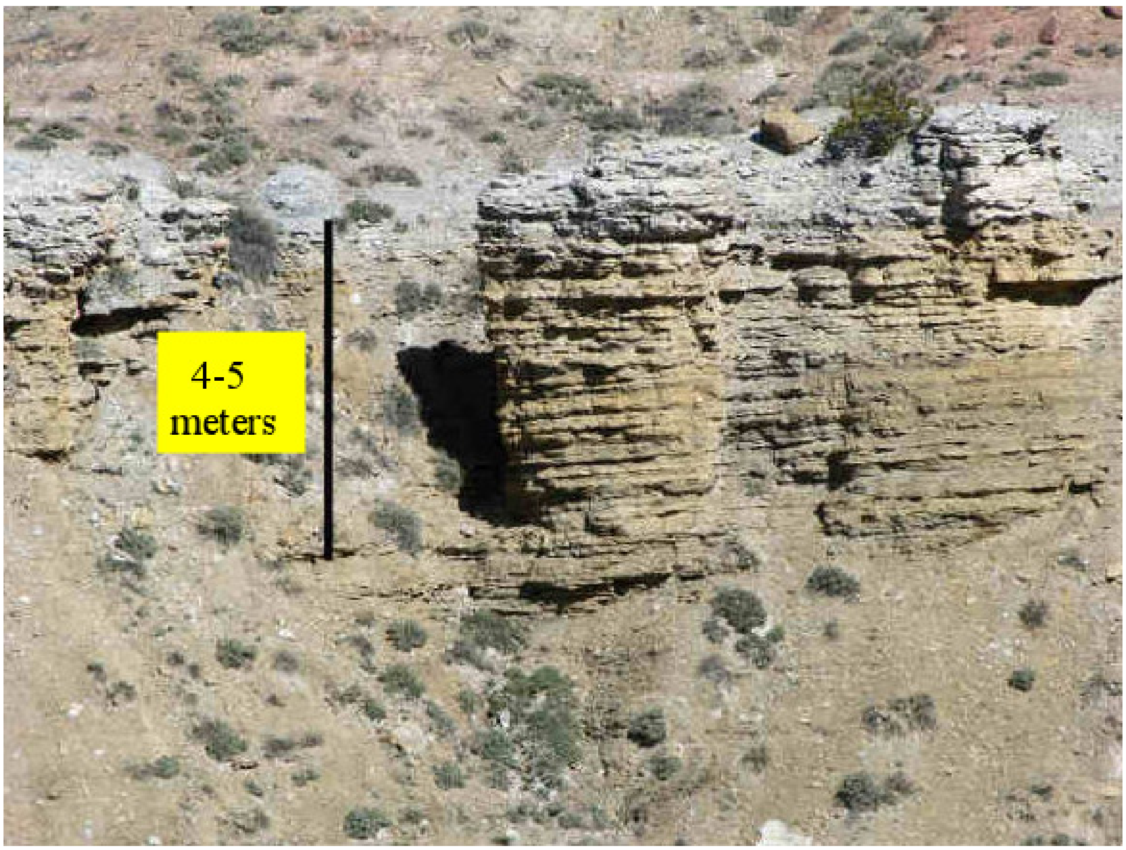



**Figure 21–These beds were observed at sedimentary log B18. The tidal flat deposits consist of centimeter-decimeter beds of sand and mud(see also Hettinger and Kirschbaum, 2002).**

Interpretation

Interbedding of thin sandstones and mudstones records rapid fluctuations in current speeds. Fine grain sizes and mostly ripple cross strata within sand beds indicate slow currents, and ripples oriented in opposite directions indicate flow reversal. Grain size variations from heterolithic sandstone and mudstone to cleaner sandstones that indicate a higher suspended-sediment concentration near the basal erosion surface. Double-mud drapes and bidirectional ripples are evidence for reversing tidal currents. Herringbone cross-stratification records reversing currents of near-equal strength. Larger flood dominated cross-stratified sandstones show increasing flow velocities. This is expected, as flood-oriented flows tend to be faster than ebb (Wright, 1985). Bioturbation in these bodies was comparable to the tidally-influenced bedsets.

**Tidal-Flat Deposits**

Description

Centimeter-thick crenulated beds of alternating mudstone to lower-fine grained sandstones typically occur in slope-forming exposures (Figures 21), which makes it very difficult to document bedding geometry. Centimeter-thick mudstone beds are massive, grey-black, and contain significant plant material. Bioturbation, ranging from 2-3, is minimal for the tidal flat deposits. Intervals of this facies are capped by a decimeter to meter-thick, laterally-continuous, sometimes white-cemented sandy bed. The top of this facies is capped by black thinly-bedded carbonaceous shale and commonly a coal.

Locally these heterolithic deposits are cut by thicker channel-form bedsets. These few meter-thick sigmoidal bedsets have erosional bases and fine upward from lower fine to upper very-fine sandstone. They are cross stratified lower down and are capped by a structureless, 2-3 meter thick clayey silt with interspersed thin sands

Interpretation

These tidal flat deposits formed within the intertidal to supratidal zone along the margins of a tidal coastline or as overflow deposits along tidal distributaries (Hettinger and Kirschbaum, 2004; Dalrymple, 1992; Figure 22). Reversing small-scale cross

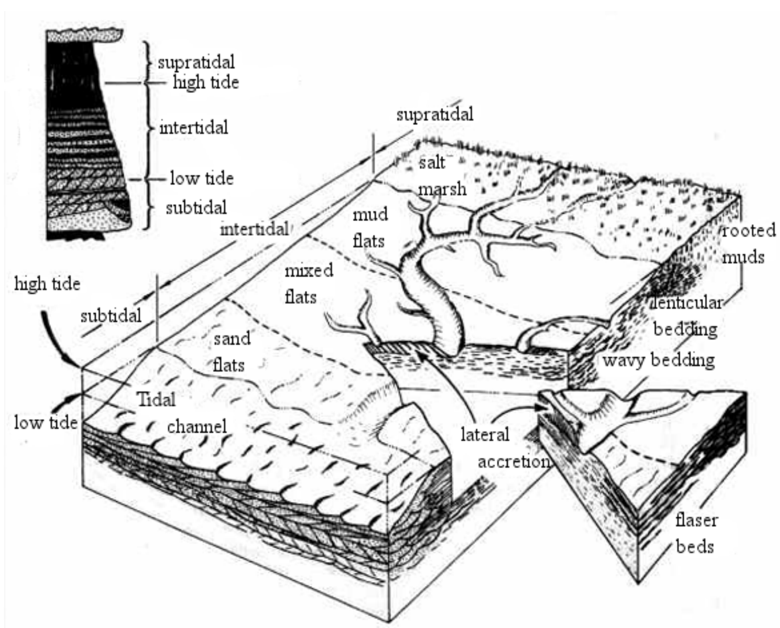



**Figure 22–This block diagram of a siliciclastic tidal flat shows the facies distribution across the region. Under the influence of the full tidal range is the sand and mixed flats. The sedimentary structures include flaser and wavy bedding. Grain size decreases up the flat as the tidal velocities flow. In the upper portion of the flat the only sands are from washover deposits that form berms on the profile (Boggs, 1995).**

stratification within interbedded, centimeter-thick sandstone and mudstone beds indicate tidal reworking. The decimeter thick benches of sand that top the tidal flat deposits are the result of storms (Devine, 1991). As tidal flats aggraded, peat deposits collected and ultimately turned to coal. Areas of the delta top stabilized by mangroves and other vegetation are in the supratidal zone (*sensu* Boggs, 1995; Allen and Chamber, 1998). Deposition of peat and mangroves greatly enhances the preservation potential of the tidal flats.

The sigmoidal–shaped sandstones are deposits of channels that migrated through the tidal flat. These upward fining successions record lateral accretion of point bars and decrease of flow velocity. Abandonment of the channel was fairly rapid, leaving a massive clayey-silt channel plug. Interspersed thin sands record small pulses of sediment as the channel was filling. The massive structure of the clay fill is related to the lack of current strength to entrain or deposit sediment. The sandy bed above this clay fill records ravinement and aggradation during continued sea-level rise.

**Cemented Layers**

Description

Although cements within the Sego Sandstone form by diagenetic processes after burial, they commonly occur in association with distinctive facies changes. Cemented zones generally are internally structureless because primary fabrics were disrupted during

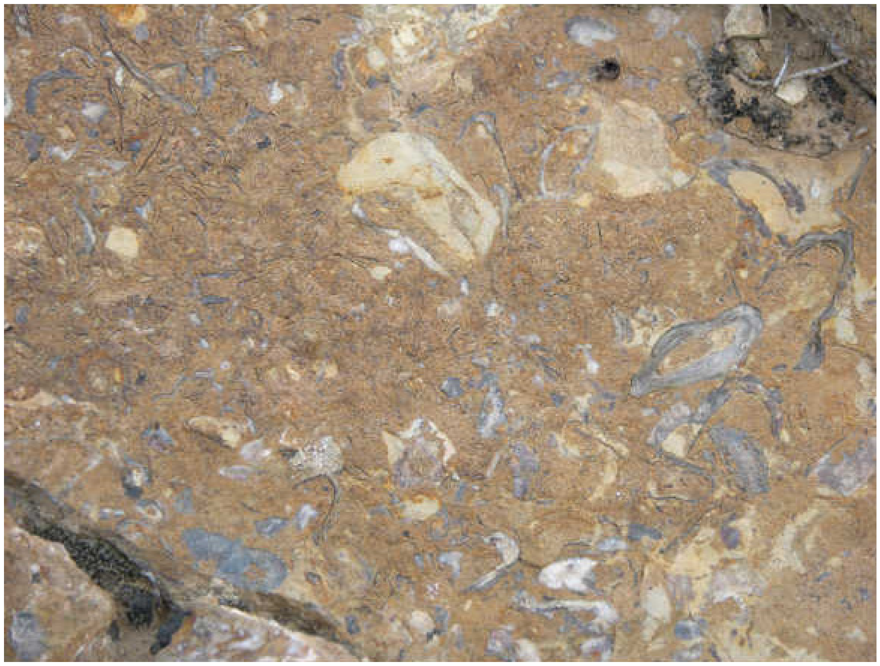


**Figure 23–Cement hash occurs throughout the upper Sego Sandstone. This oyster rich unit is interpreted to imply the transgression of oyster shoals in the intertidal environment.**

the cementing process. Three types of diagenetic beds are defined: regionally traceable cements that commonly have oyster lags, discontinuous cemented layers, and white-capped cemented sand benches.

Regionally continuous layers of ferroan dolomitic cement are pale to grayish red and commonly occur where thicker sandstone abruptly fine to shale. In many cases a discontinuous fossil hash is associated with these cements, composed of a mixture of fossils dominated by unbroken oyster shells (Figure 23). Cemented layers tend to be thicker where grain sizes are coarser.

Discontinuous cemented zones are a few meters thick, and meters to several hundred meters wide. Discontinuous cements do not follow depositional layering. Cements are commonly associated with intervals highly bioturbated by *Ophiomorpha*, *Planolites*, *Chondrites*, *Paleophycus*. They also occur in less bioturbated deposits associated with localized lags of whole or incomplete oyster shells.

White leached benches generally underlie deposits with high organic content, like the tidal flat facies with coals. Viewable within limited areas these zones are traceable across poorly exposed areas as a scattering of whitish-green chips (Figure 24).

**Figure 24–White cemented benches occur below carbonaceous shale and coal layers. These layers do not cross sequence boundaries.**

Interpretation

Although cements are common throughout the Sego sandstone, their concentration along particular horizons and within different zones can be related to depositional shell concentrations and pathways of subsurface fluid movement through the deposits. Thus, they commonly highlight specific types of allostratigraphic surfaces and times when marine ground waters moved through specific depositional facies.

Continuous cements associated with discontinuous oyster-bed hash may form during high-frequency sea-level rises, when tidal action pumped marine waters through permeable sediments. This mixing of phreatic or meteoric waters in an interval with abundant organic-rich mudstones precipitated cement within the strata. Organic material above the sand provides a source of bicarbonate resulting in iron reduction (Taylor et al., 2000). These diagenetic processes may be enhanced by decreased sediment input during relative rises in sea level.

Discontinuous cements reflect early diagenetic conditions as meteoric waters move into more distal areas during relative sea-level falls. Taylor et al. (2000) suggest that in order to get these large-scale cemented zones extreme subsurface flow rates are required. Other sources for marine carbonates come from fossils such as oysters and detrital dolomite. This was supported in other parts of the Book Cliffs by cemented bodies being found in shoreface sandstone down-dip from high frequency sequence boundaries in the Grassy highstand sequence (Taylor, et al., 2000).

White capped sand-bodies are distinctive and easy to trace from a distance. They result from the leaching of detrital dolomite under coal deposits. A flux of acidic fluids appears to have been restricted to less than 5 meters below a coal layer. Coals are not observed within the upper Sego Sandstone here, but are reported to be preserved within this interval further east in Colorado (Kirschbaum and Hettinger, 2004).

# STRATAL ARCHITECTURE

The architecture of the upper Sego sandstone is defined by organized vertical variations in facies and continuous allostratigraphic surfaces. Three types of stratigraphic surfaces are recognized: 1) Erosion surfaces that define an abrupt coarsening from shales to sandstones; 2) high-relief erosion surfaces within sandstone layers; and 3) surfaces defined by cements or an abrupt fining from sandstone to shale. These surfaces and facies trends were used to divide the upper Sego sandstone into two intervals. Variations that define these intervals are described and interpreted below, before the broader evolution of depositional environments and facies preservation are discussed within a sequence stratigraphic framework.

## Description

The Anchor Mine Tongue is the marine shale with wavy sandstones. There are two, meter-thick, upward-coarsening successions within the upper 2 meters of this shale. Thin beds of hummocky cross-stratified sandstone caps the upper succession in the western part of the cross-section (Figure 25).

The base of the upper Sego is an erosion surface abruptly overlain by upper very fine sandstone. The Basal layer within the upper Sego deposits comprise two bedsets of highly bioturbated lower shoreface sandstones, separated by a cemented layer. The lower bedset is up to 7 meters thick and the upper is up to 5 meters thick. Both bedsets are thickest in the eastern part of the study area (Figure 11, logs S1 and S2), and thin markedly to the west (Figure 12, log sites B8-B20). The top of the second bedset, like the

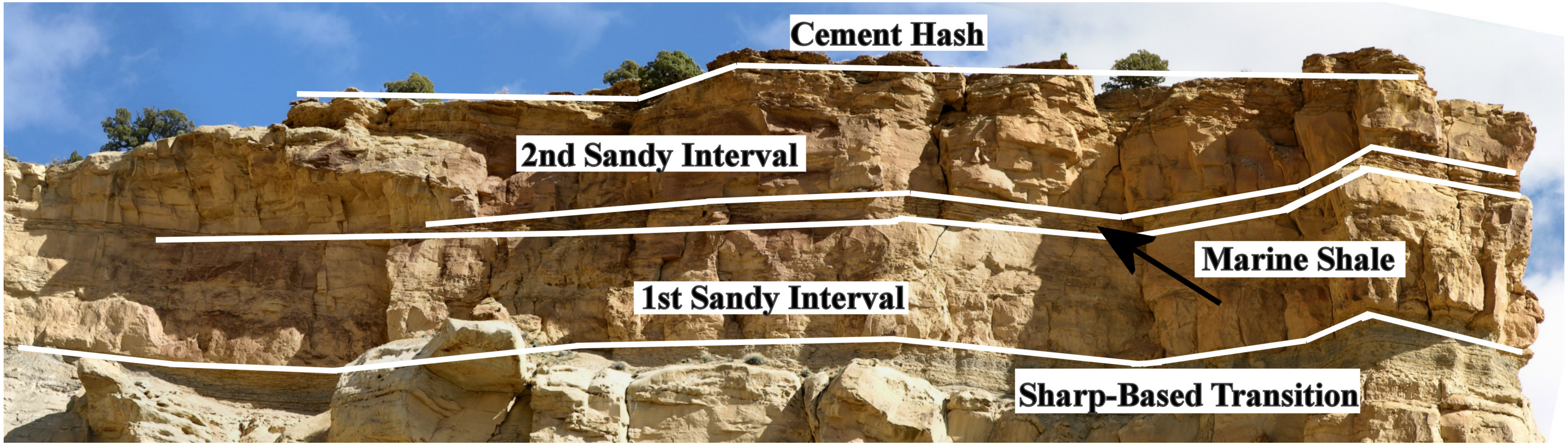



**Figure 25-This outcrop photomosaic from log B13 may be used as a reference for the generalized stacking pattern of the two sandy intervals found within the upper Sego Sandstone. The sharp-based lower shoreface is common to the majority of the outcrop. The marine shale is commonly replaced with a ravinement surface and or a cement.**

first, is distinguished regionally by well-developed cement layer, except where it is cut out by an overlying erosion surfaces between logs B14-B18. Although continuous and locally nearly a meter thick, these cements are not associated with fossil lags. Discontinuous cements occur locally throughout both bedsets.

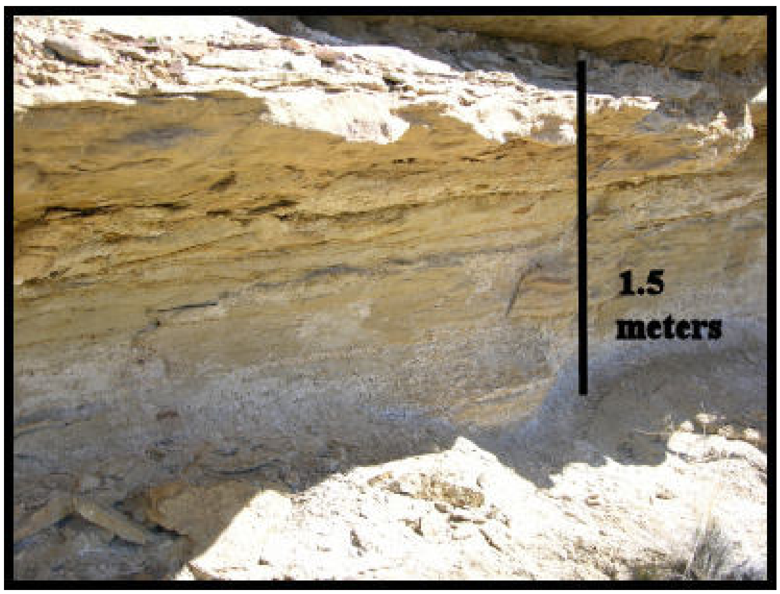



**Figure 26–This photo shows the heterolithic nature of the upper tidal bedsets. Note the easily eroded lower part of the sandstone and the substantial coarse sand above.**

An erosion surface cut into the basal bioturbated bedsets is defined by an abrupt transition to tidally influenced shoreface bar deposits. In the eastern part of the cross section these tidal bar deposits are less than 2 meters thick and are extensively bioturbated by *Ophiomorpha*. Towards the western end of the cross section these sandstones thicken and comprise distinct stacked bedsets that shingle to the west. One bedset thins to the west and passes laterally into muddier wave-influenced deposits before an overlying bedset thickens more than 10 meters and cuts it out between logs B17 and B18. The overlying bedset then gradually thins farther west, before it is in turn is cut out by the steep erosional margin of another bedsets at the western edge of the cross section. The top of this tidal sandstone layer is more bioturbated, particularly by *Ophiomorpha*. These bioturbated sandstones are overlain abruptly by a meter-thick layer of marine shale.

An erosion surface defined by an abrupt coarsening to fine sandstone cuts out the shale layer in the western part of the cross section, and it is defined by a continuous cement bed where it incises into the tidal sandstone layer below. Above this erosion surface is another 10 meter thick layer of stacked sandy tide-influenced shoreface bar deposits. Three stacked bedsets in the eastern end of the cross-section are each 7-8 meters thick and shingle to the west. Bedsets are thicker and sandier on average than those in the lowerpart of the cross section. Some examples become more heterolithic as they shingle toward the west. Bioturbation is sparse and does not increase towards the top of each bar. Bedsets within the upper 5 to 8 meters of this layer are more heterolithic than those below along the western end of the cross section, but not in the eastern part (Figure 26). Bedsets are separated by continuous cements. A particularly thick cemented bed at the top of this stack of tidal bedsets is associated with fossil oyster hash.

A high-relief erosion surface locally cuts down from the capping fossil bed through the layers of stacked tidal bedsets just described; in places cutting downward nearly to the Anchor Mine Tongue (Figure 27). Relief along this erosion surface defines several distinct incisions, each about a kilometer wide. Deposits that fill these incisions are more heterolithic than tidally influenced bedsets below the erosion surface, and in some cases these deposits are dominated by mudstone. Although coarser grained incision fills show only subtle vertical facies trends, finer grain fills display pronounced upward-coarsening trends. For example the fill documented in Log C2 is dominated by mudstone with thin wavy sandstone beds lower down, and grades into sandy cross stratified sandstone upward. The sandier incision fill documented in log B10 contains thinner beds lower and thicker beds higher in the fill, but it is dominated by sandstone throughout. Poor exposure of these fine grained facies limits documentation of bedding geometry, but cement layers are commonly continuous across incision fill even in areas where lithic beds are difficult to trace. The pronounced cement bed and associated fossil hash layer that caps upper Sego sandstone appear to extend also across deposits incised by this erosion surface.

Eight meters of very heterolithic tidal flat deposits overlie the shell hash layer that caps the upper Sego Sandstone. Because these beds are deeply weathered, lithic variation could only be observed locally. The basal deposits are dominated by shale within heterolithic layers richer in thin bedded sandstone. Locally, small heterolithic channel-form bedsets cut into these heterolithic deposits. The interval is capped by a thicker, very

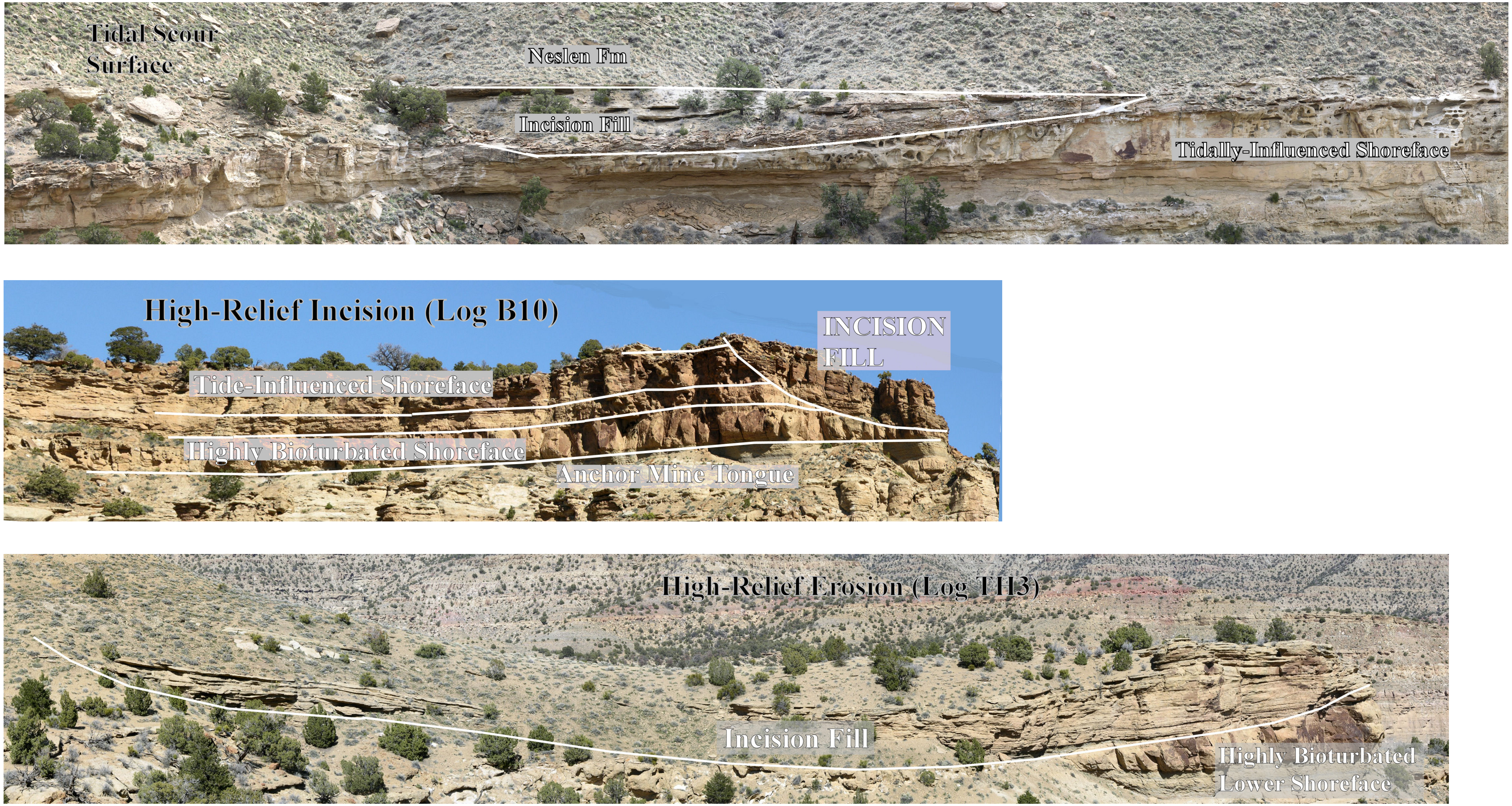



**Figure 27-Distributary channels observed in the upper Sego Sandstone show scoured bases and tidal bedding. The channel shapes tend to be symmetrical and have high width/thickness ratios. Beds coarsen upward and do not exhibit any fluvial facies.**

bioturbated sandstone bench overlain by a thin carbonaceous shale and coal (e.g., Figure 12, log B20).

**Interpretation**

Shales of the Anchor Mine Tongue record regional transgression onto the lower Sego Sandstone (Young, 1966; Kirschbaum and Hettinger, 2004). Interbedded shales and thin wavy sandstones within the Anchor Mine Tongue were deposited between storm and fair weather wave base in fully marine waters. The highly bioturbated lower shoreface sandstones that directly overlie the basal erosion surface of the upper Sego Sandstone reflect relatively slow deposition on the lower Shoreface. Typically deposits formed during the regression of shorelines are expected to progress gradually upward from offshore prodeltaic mud to heterolithic mud and sands and finally to clean sandy deposits of the shoreface (Reading, 1982). The abrupt vertical transition of these deposits at the base of the upper Sego Sandstone suggests a stratigraphic discontinuity. Many studies have suggested that similar discontinuities can form by submarine erosion, where shorelines regress due to falls in sea-level that lower wave base onto more distal areas of the sea floor (see Plint 1988, and reviews in Plint and Nummedal, 2000; and Posamentier and Morris, 2000). Although the lack of primary sedimentary structures within these sandstones makes it difficult to know whether these sands where deposited under waves or tides, flat erosion surfaces like that seen here are more likely to be formed by wave erosion because wave base tends to be constant over large areas. The thinning of deposits to the west suggests these sands originated from a point source (e.g., the distal end of a lobe extending away from a distributary rather than a lineal wave deposited barrier sand).

Tide-influenced shoreface bars above these bioturbated lower shoreface deposits occur abruptly across a distinct erosion surface, and thus this facies change from the highly bioturbated sandstone below indicates another stratigraphic discontinuity that marks an increase in tidal current strength and deposition rate. Distinct bars and more heterolithic facies suggest deposition closer to the shoreline. These deposits thus may record continued regression during sea level fall and erosion of the lower shoreface deposits by stronger tidal currents near shore. Alternatively, the continuous erosion surface at the base of this facies may indicate a period of lowstand erosion and bypass of sediment farther basinward, and the tidal facies above this erosion surface may have formed during transgressive reworking of lowstand fluvial and shoreface deposits during transgression. In either case, the bioturbated meter-thick cap of this sandstone interval and the overlying shale records transgression and flooding of the shoreline.

The erosion of the transgressive marine shale in the middle of the upper Sego Sandstone by overlying tide-influenced bedsets records renewed progradation of the shoreline. Although this erosion surface has more relief than the one at the base of the upper Sego Sandstone, it probably formed by similar processes of marine erosion. The lack of lower shoreface deposits in the lower part of this interval suggests significant erosion by shoreface tidal currents and perhaps less rapid accumulation of lower shoreface deposits during regression. Erosion along this surface was greater to the west, where it incises through the shale layer into the tidal sandstone below. Shingling and truncation of successive bedsets within this layer indicates a progressive shift in sediment accumulation to the west. More heterolithic bar deposits near the top of the upper Sego Sandstone along the western part of the outcrop may record weaker currents, higher suspended sediment concentrations, and lower deposition rates than sandier bedsets along the same interval to the east. Cement or oyster shell hash capping individual bedsets records the development of oyster shoals after tidal bar abandonment; presumably associated with avulsion of a distributary.

Interpretation of the high-relief erosion surfaces that descend from the top of the upper Sego Sandstone is problematic; these are either basal erosion surfaces of tidally enlarged distributary channels or they record multiple locations of lowstand incision. These deposits are similar in scale to tidally-enlarged, distributary channels cut into an abandoned area along the Fly River delta front (Dalrymple, 1997), and similar channels cut off from the fluvial source along the modern Mahakam Delta (Allen and Chambers, 1998). These heterolithic fills presumably then reflect the subsequent filling of these incisions by tide reworked sediment carried along the coast from areas of more active sediment progradation. The alternative, that these surfaces can be linked up to define a major stratigraphic discontinuity, which would indicate there was fluvial deposits in the base of these incisions that were removed during subsequent transgression.

The thick cement and fossil lag that caps the upper sego sandstone records more regional ravinement during transgression. Tidal flat facies above this erosion surface accumulated as seas regressed and then vertically stacked, heralding the beginning of Neslen Formation deposition. The channel found at log C2 with upward fining heterolithic fill is a tidal channel deposit. The mud filled channel formed when the

**Figure 28–Fluvial channel fills in the area of b20 fine upward from laterally accreted surfaces to predominantly muddy grey channel fills. This fill is generally accepted to be formed from the avulsion of this minor channel to another source.**

distributary channel was shielded during renewed progradation and filling (Willis and Gabel, 2003). To the far west of the survey area (log B20), a channel fill with internal lateral accretion surfaces is eroded into the tidal flat facies (Figure 28). Moderate to high sinuosity (meandering) channels are characteristic of tidal channels far inland from the shoreline (Figure 29).

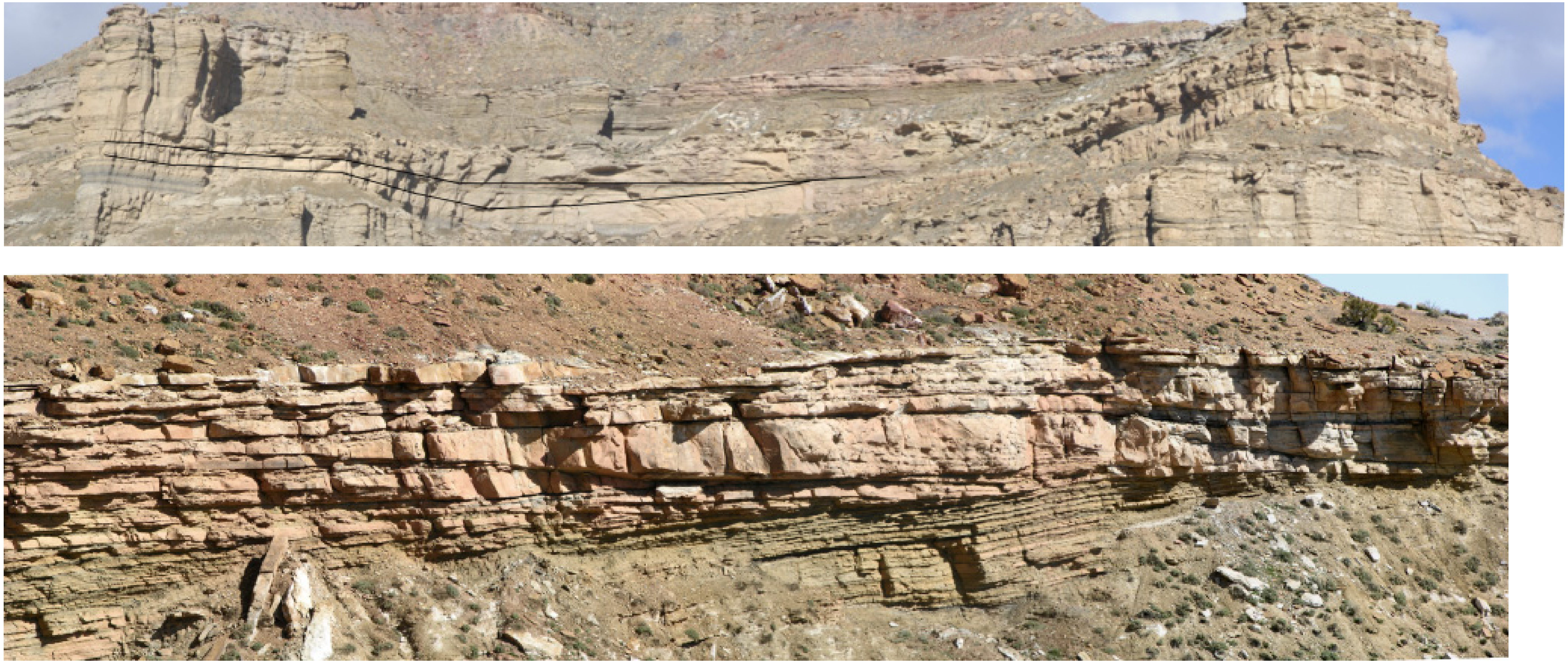


**Figure 29-The first photomosaic displays the unmeasured incision west of log B20. The fill of this channel appears to fine upward but was not safely accessible. The second photomosaic displays the channel measured ar log B20. This minor channel is one of the most recognizable features in this area of the Book Cliffs. Laterally accreted bedforms show the migration of the channel as it incised the tidal flat facies.**

The upper Sego Sandstone is divided into two sandy intervals separated by laterally continuous shale. Each interval has an erosional base overlain by deposits interpreted to have formed on a prograding shoreface. The first interval contains a thick basal succession of lower shoreface facies. The second interval, thicker than the first, contains only upper shoreface deposits. The overall succession is interpreted to be forward stepping; deposits within successive intervals recording on average shallower-water deposition. This broad interpretation is complicated by two issues 1) the dominance of flood oriented paleocurrents within the tidal bar facies may suggest deposition during transgression, 2) the high-relief erosion surfaces within sandy intervals can be interpreted to reflect different types of discontinuities. These two issues are discussed below, before sequence stratigraphic interpretations are addressed.

## FLOOD-ORIENTED DEPOSITS WITHIN DELTAIC STRATA

Sediment transportation patterns on tide-influenced river deltas are complicated by the interaction of ebb-dominated river flood currents and the common dominance of flood sediment transport (Figure 30). Since river floods occur rarely relative to the frequency of tidal cycles, observed paleocurrent indicators may not reflect net movement of sediment from distributaries into the basin. Interpretation of upper Sego Sandstone deposition is problematic because vertical facies trends suggest long-term regression whereas bedset accretion and smaller scale

**Figure 30–This flood oriented bar is found in relation to the shallow tidal scour found within Blaze Canyon. These slightly dipping features are common to the upper Sego Sandstone.**

paleocurrents indicators suggest broadly flood-dominated sediment transport directions. Although there are several studies that report patterns of sediment transport within tide-influenced estuaries (Coleman et al., 1988; Dalrymple and Rhodes, 1995; Harris et al., 2004), there are relatively few modern analogues that can be used to predict preservation

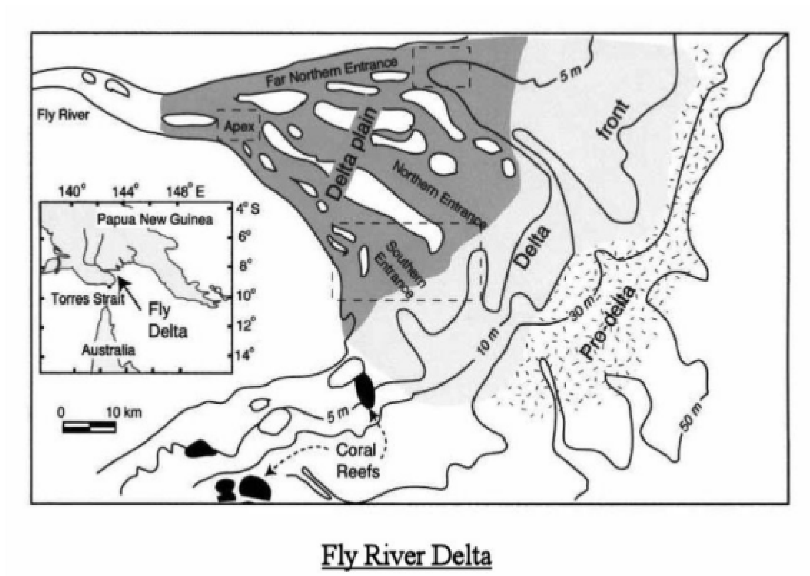



**Figure 31–Fly River Delta in Papua New Guinea is the closest approximation to a modern analogue of the upper Sego Sandstone (Harris et al., 2004).**

of flood transported sediment on prograding tide influenced deltas. Important insights from estuarine systems are that ebb and flood tidal currents tend to move along mutually-evasive pathways onshore and offshore and that the orientation of flow indicators within preserved sediments depends more on where sediments accumulate within these systems than on which direction of currents dominate overall. Thus where the faster direction of sediment transport is associated with erosion and the direction of slower currents leads to deposition, flow indicators in the subordinate direction will be preferentially preserved.

The Fly River delta, Papua New Guinea, is one of the few tide-influenced deltaic systems where sediment transport processes and net sediment transport budgets have been documented (Dalrymple et al., 2003; Harris et al., 2004; Figure 31). Current measurement in distributary channels over a month period indicated that individual distributaries were variably ebb or flood dominated. Dominant flow directions were not, however, directly related to directions of net sediment transport. During large spring tides, there was no net sediment transport. During neap tides, net transport was landward. An ebb tide carries an exponentially lower suspended sediment concentration during. (Geyer et al., 2001). Intrusion of a salt wedge during weak neap tide and increase in density stratification as fresh water flows out as a hypopycnal flow tended to increase preservation of flood-oriented, small-scale bedforms.

Normal river discharge and ebb tidal currents did not appear strong enough to carry sediments offshore, because flood tide currents are accentuated by the funnel shape of the delta and shoaling into the delta front. The rapid mixing of fresh and salt water along the delta front can lead to rapid flocculation of clays, which may then move down slope as fluidized mud. Infrequent major river floods presumably would transport sediments basinward, but these currents may tend to erode delta top and delta front areas as sediment is bypassed further basinward. Salt-wedge intrusions into the distributary channels would be expelled during river floods and with it an increase in suspended sediment concentrations (Geyer et al., 2001). River currents rising in a hypopycnal plume during neap tides would no longer produce ebb-oriented bedforms. Following a river flood, tidal currents may rework sediment landward onto the delta front, preferentially preserving flood-oriented tidal currents (Woodruff et al., 2001). Flood-tide current velocities may exceed that of ebb currents in most tide-influenced systems, resulting in a net movement of sands shoreward (e.g., Ord River, Wright, 1985). Wright (1985) also reports that the deepest channels on deltas, cut by ebb flows, tend to fill after distributary channel abandonment with flood-tide transported sediment.

The migration of tidal bars may preferentially preserve smaller-scale cross stratification oriented in the subordinate current direction (Dalrymple, 1992). In the macrotidal Han River Delta in Korea, a detailed survey of inclined heterolithic stratification within a large tidal channel point bar showed that seaward-dipping accretion beds were dominated by flood-oriented internal cross-strata (Choi et al., 2004). Ebb currents dominantly flowed along the channel cut bank side where sediments are being eroded. Flood currents decelerate onto the basinward end of the point bar gradually extending the bar. Tidal rhythmite deposition during flood tidal flows on the basinward ends of bars was enhanced in narrower, higher-sinuosity channels, where less erosion occurred along the downstream ends of bars during subsequent ebbs flows.

The reworking of sediment supplied by river floods landward by tidal currents during periods of normal river discharge is supported by studies of dune formation within the Gironde Estuary (Berne et al., 1993). The tidal bedsets observed reversed their asymmetry during minor changes in river discharge. In tidal channels, 5 km long and 1 km wide, dunes changed their shape and orientation over just a few days (as few as 19 tidal cycles). Ebb-oriented dunes were reoriented by the weaker neap tides into flood-oriented dunes over longer periods 96 days, 185 cycles).

Larger-scale patterns of sediment transport along prograding tide-influenced deltas may also preferentially preserve flood oriented deposits in local areas. Where tidal waves move onshore obliquely to the coast, there will be net along shore sediment transport

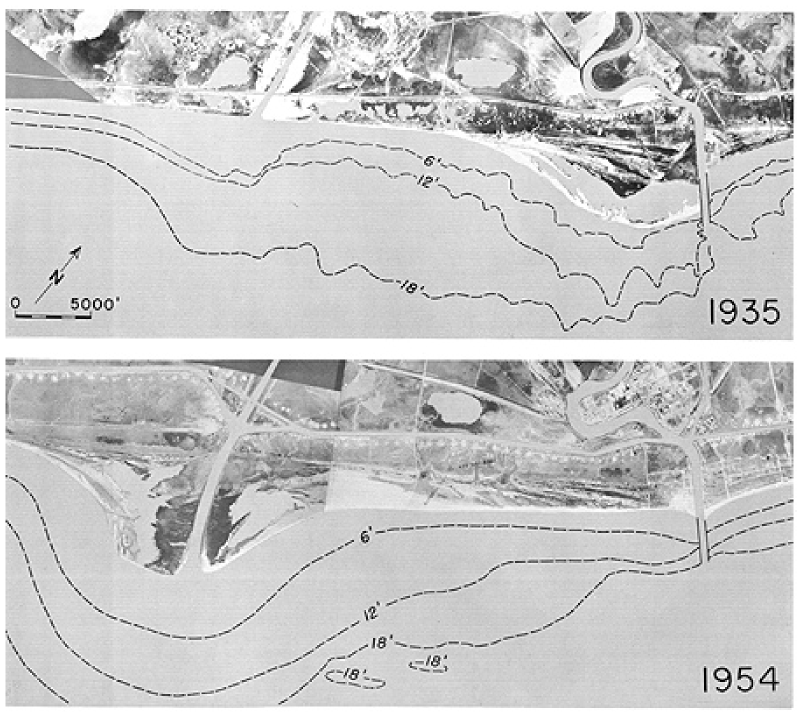



**Figure 32–The Brazos River Delta was extensively reworked after its abandonment in 1929. This bathymetric map shows the extent and speed of the lobes erosion (Bernard et al., 1999).**

(Wright, 1985). As river supplied sediment builds into the basin, tidal currents may preferentially rework sediment onshore into adjacent embayed areas. This reworking of deltaic sediment may be particularly pronounced after distributary avulsion abandoned a delta lobe, providing more time for tidal reworking. Evidence of this would be bedsets showing basal ebb-dominated flow-orientated deposits, internal erosional scours through the lower to central portions, and upper flood-dominated flow-orientations deposits. Deposits can be reworked very quickly as large portions of the lobe are lost to erosional currents. The uneven topography of the paleo-coastline provided the accommodation space necessary for the emplacement of several meter-thick bars of sediment.

Although examples of the tidal reworking of abandoned delta lobes are not well documented, the extent of reworking possible is documented for wave-dominated deltaic systems. The wave-dominated Brazos River Delta along the northern Gulf of Mexico coast was diverted in 1929 to a new area (Rodriguez et al., 2000; Figure 32). By 1931 the subaerial portion of the abandoned delta lobe had been reworked to wave base. Currently (<75 years after the diversion) the entire protruding delta has been eroded and transported down-coast. Sediments of the Amazon River Delta are reworked by Guiana marine current several hundred kilometers from the delta front along the coast (Wright, 1985). The Po River Delta has been extensively reworked by wind-driven currents during storms (Fox et al., 2004).

Interpretation of paleocurrent orientations within the Sego sandstone require a better understanding of sediment transport and preservation within tide influenced systems. It may be common in tide-influenced deltaic systems for flood tides to rework sediments onshore following river floods. In a broad shallow sea, like the Western Interior Seaway, tidal currents may have significantly reworked river supplied sediment along the coastline. In prograding wave-dominated systems, a distinction is common between delta deposits (where sediment bodies reflect deposition away from the distributaries that supplied sediment) and strand plains formed where waves have completely reworked river supplied sediments along the coast. Perhaps a similar distinction should be advanced for tide-influenced deltas and tide-dominated systems where sediment is reworked far from river mouths. Because of the difficultly in relating paleocurrent indicators to directions of net sediment aggradation, it is probably better to interpret regional regressive and transgressive trends based on broader changes in facies.

## ORIGIN OF HIGH-RELIEF EROSION SURFACES

High-relief channel-form incisions cut through prograding clastic shoreline deposits have traditionally been interpreted to be either distributary channels or lowstand valley fills. The distinction is important because distributary channels record only local sediment bypass to the delta front whereas valley fills suggest more regional sediment bypass to broader systems of lowstand deposits. The distinction is particularly difficult in shallow-sea ramp settings, like the Cretaceous Western Interior Seaway, because sea level falls do not expose a tectonic shelf break that can provide knickpoints for deep valley incision. Rather valleys are expected to incise only as deeply as the thickness of stranded shorelines in more proximal areas of the basin. Several criteria have been proposed to distinguish distributary channel from more regional valley incision. 1) Valleys occur along regional erosional surfaces and are larger than a single channel. 2) Since distributary channels feed associated shoreline deposits, they generally do not cut completely through laterally adjacent upward-coarsening shoreline deposits. 3) Since bars scale to the full discharge of a channel, bedsets that fill channels should be the same scale as the incision (or at least fill a significant part of the incision; 80%). Valleys, in contrast, should contain multiple vertically amalgamated channel bar deposits or beds of estuarine deposits that onlap the valley wall. 4) Regional erosion surfaces associated with valley incision may be traceable across adjacent interfluves as paleosols, pebble lags or *Glossifungites* ichnofacies. 5) Tributaries tend to incise into valleys but not at junctions with another channel (Plint and Wadsworth, 2003).

High relief erosion surfaces that incise into the two sandy layers of the upper Sego Sandstone documented here have some aspects similar to those used to define valleys: both can be traced continuously across the outcrop, deposits above these erosion surfaces clearly comprise either multiple, vertically stacked bar deposits or more heterolithic estuarine deposits that onlap the incision margin, and the upper of these erosion surfaces cuts deeper than the thickness of the adjacent shoreline deposits. There are, however, several observations that argue against the interpretation that these erosion surfaces are formed by lowstand fluvial incision. The incision through the lower sandier interval has relief not much greater than tthe thickness of an individual tidal bedsets; relief similar to the surface at the base of the second sandy interval interpreted to reflect falling-stage tidal erosion. Although the second surface has significantly greater relief, individual incisions along this surface are less than a kilometer wide; similar to distributary channel deposits observed elsewhere within the Sego Sandstone. Incision fills above this upper erosion surface commonly coarsen upward, in contrast to the fining-upward estuarine deposits expected to be deposited during the transgression of a valley.

An alternative explanation is that deep incisions into the Sego Sandstone record enlargement of distributary channels by tidal erosion (Willis and Gabel, 2003). Such enlargement may be more common after an area along the delta has been abandoned by distributary channel avulsion, for then tidal-scouring would have more time to deeply erode the delta front. Several modern analogues suggest this process is likely to occur. When the major discharge from the Fly River avulsed from the northern to southern part of its delta, tides enlarged the northern distributaries from 5 meters to almost 30 meters deep in a period of only a 100 years (Dalrymple, 1997). Unlike valleys, distributaries can be abandoned and reoccupied due to local shifts in sedimentation on the delta, and thus need not be associated with regional shoreline regressions and transgressions. Reoccupied distributaries may be filled with prograding deltaic deposits or by transgressive successions formed as sediment is transported along shore away from areas of more rapid sediment input along the shoreline.

## SEQUENCE STRATIGRAPHY

Sequence stratigraphic interpretations of the Sego Sandstone vary widely: Ranging from the suggestion that nearly all these deposits formed during regional transgression within a succession of incised valleys formed during a succession of high frequency sea-level fluctuations (Van Wagoner et al., 1991; Willis, 2000), to the idea that these deposits record episodic progradation, abandonment, and tidal erosion of deltas during a regional regression (Willis and Gabel, 2003). The contrasting sequence stratigraphic models have very different implication of preservation of tide-influenced depositional environments and the volume of sediment bypassed farther into the basin. The key to distinguishing these different interpretations is an understanding of the significance of stratal discontinuities.

The first sequence stratigraphic models of the Sego Sandstone (Van Wagoner et al., 1991) assumed that any abrupt vertical change from finer-grained, deeper-water facies to coarser-grain, shallower-water facies or incision records an abrupt forward shift in deposition due to a lowstand in sea level. Using this idea the upper Sego sandstone documented in this study would comprise 4 high-frequency sequences, each bounded by an erosion surface (i.e., the erosion surface at the base of the upper and lower sandy intervals, and erosion surfaces within these sandy intervals). Each of these erosion surfaces would be interpreted to record lowstand sediment bypass to more distal areas of the basin. Van Wagoner et al. (1991) suggested that sea level falls shifted deposition basinward so fast that few regressive (highstand) sediments were deposited and those that did form were uniformly reworked into incised valleys during transgression. The upper Sego Sandstone deposits documented here would thus nearly all be lowstand valley fills, because both falling-stage highstand and transgressive deposits were eroded landward during sea-level rise. These sequence boundaries would be related to fourth-order sea-level cycles (Van Wagoner et al., 1991). These cycles are related to changes in sea-level initiated by tectonic events in the Sevier Orogeny (Willis, 2000). The relative time range for this fourth order change in sea-level is 100,000 to 1,000,000 years (Krystinik, 1995).

More recent sequence stratigraphic models are based on the idea that significant erosion along tide-influenced shorelines can form by autocyclic processes (Figure 33; Willis and Gabel, 2003). It was observed that erosion surfaces at the base of sandy intervals were not regionally continuous and that stratigraphy defined by criteria of Van Wagoner et al. (1991) resulted in an ever growing number of sequences as they were traced over broader regions. Recently an increasing number of stratigraphic studies of wave-dominated systems interpreted erosion surfaces at the base of prograding shoreline deposits to record scouring of offshore muds as sea levels fell (i.e., regressive surface of marine erosion; Plint 1988; Plint, 2003; reviews in Plint and Nummedal, 2000; and Posamentier and Morris, 2000). Studies of deposition along modern tidal shorelines suggest marine erosion and sediment transport by tides is common at a variety of water depths, and may be enhanced as seas shallow during forced regression. Based on these ideas, erosion surfaces at the base of the two sandy intervals documented within the upper

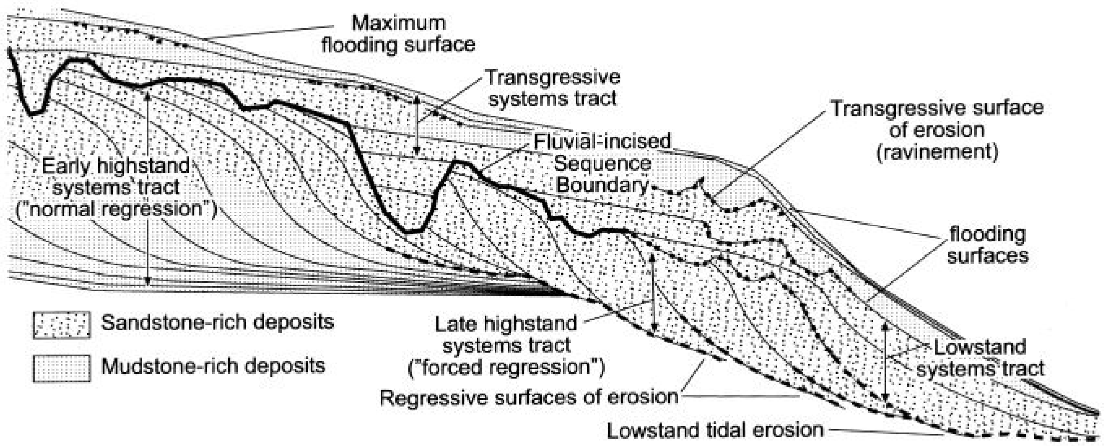



**Figure 33–This sequence stratigraphic diagram illustrates interpretations for the lower and upper Sego Sandstones. Falling-stage/late highstand systems during renewed progradation and falling sea level will incise the sea floor causing sharp-based deposits to form (Willis and Gabel, 2001).**

Sego Sandstone are interpreted to record increased tidal erosion and a basinward shift in facies during falls in sea level (Figure 34). The erosion surface through the lower sandstone interval has similar relief and may be a similar regressive surface of marine erosion. Deposits preserved within the upper Sego Sandstone are thus interpreted to be all late highstand, formed during shoreline regression. The shale bed separating lower and upper sandy intervals is interpreted to record a short term transgression or relative sea-level rise, likely associated with local distributary avulsions. The only candidate for a valley (Exxon "sequence boundary") is the high-relief erosion surface that cuts down from the top of the upper Sego Sandstone, but these incision depths can alternatively reasonably be associated with tidal enlargement of tidal channels. These incisions may correlate laterally to a lowstand surface of erosion (sequence boundary) filled with

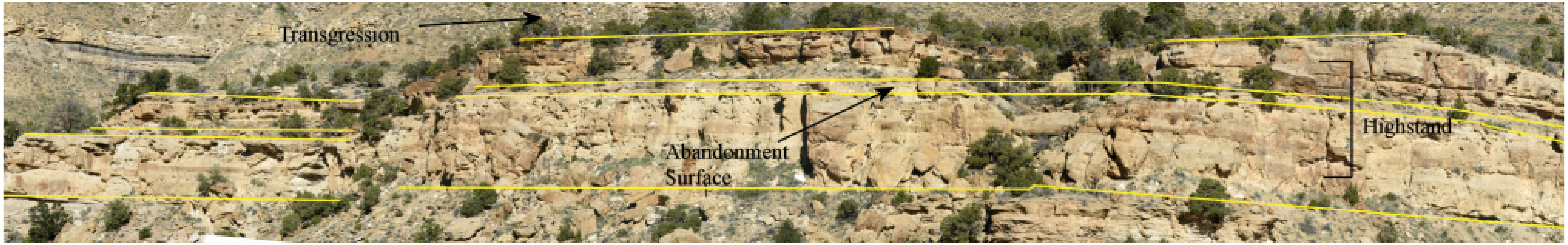



**Figure 34-The photomosaic of log T3 in the eastern part of the survey shows one of the possible sequence stratigraphic interpretations for the upper Sego Sandstone. Note the separation between sand intervals. During this period deposition was induced by distributary abandonment.**

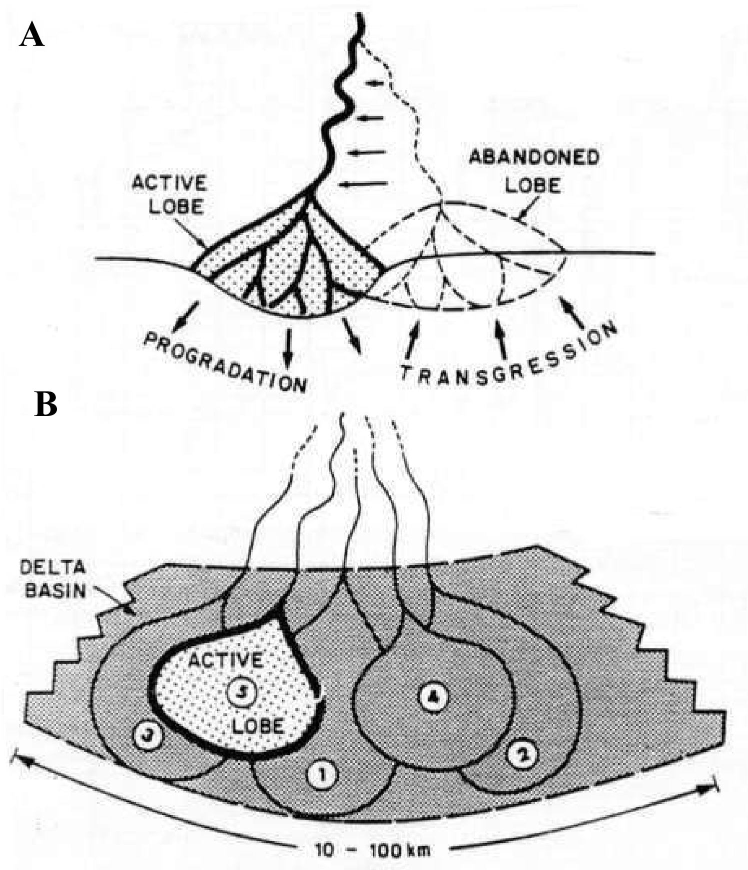



**Figure 35–The abandonment of deltaic lobes is a common occurrence when delta distributary mouth bars aggrade too high and limits the flow of the alluvial system. A) Fluvial waters will reroute upstream in order to gain the gradient advantage and find the most expedient way into the basin. B) This can also occur during times of sea-level lowstand when waters recede and highstand deltas are incised to deposit a new lobe towards the toe of the previous lobe (Allen and Chambers, 1998).**

amalgamated channels observed to cut through the upper Sego Sandstone into the lower Sego Sandstone in outcrops west of the area documented here. Incisions may either reflect valley segments that were not occupied by fluvial rivers during transgression or they may have formed by transgressive tidal erosion into the maximum regression surface during transgression. Alternatively, the lowstand surface of erosion observed to the west of this study area may correlate with the basal tidal flat deposits found well above this incision surface. This would associate the high-relief incision and its upward-coarsening fill to a subsequent late-highstand. This highstand systems tract must then have been terminated by the avulsion of rivers westward, which fed the basin and coastline away from this lobe allowing the aggradation of transgressive tidal flats (Figure 35). Lack of appreciable eustatic sea-level changes during the Upper Campanian (Franczyk, 1989) supports the interpretation of a highstand systems tract without major lowstand surfaces of erosion. Given the coastal ramp geometry of the seaway into which the Sego Sandstone prograded, tectonic elements of the foreland basin, and the limited extent of this study; current sequence stratigraphic nomenclatures are not well suited to describe these sandstones (Ryer, 1993).

## CONCLUSIONS

The upper Sego Sandstone provides insights into preserved facies of tide-dominated deltas. It is a 20-30 meter thick succession that extends from east-central Utah into Colorado. Stratigraphic logs measured along an 8 km long outcrop and photomosaics of several extensive cliff exposures were used to construct an oblique to strike-oriented bedding diagram showing lateral and vertical facies variations and the geometry of stratigraphic surfaces. The following conclusions were drawn from analysis of these observations.

The documented deposits within the upper Sego Sandstone through basal Neslen Formation interval comprise five distinct facies types: marine shale with wavy sandstones, highly-bioturbated lower shoreface, tide-influenced shoreface bars, heterolithic tide-influenced shoreface bars, and tidal-flat deposits.

Vertical progression of facies within the upper Sego Sandstone records the episodic progradation of tide-influenced river deltas during falling sea level. Continuous erosion surfaces at the base of and within the first sandstone interval record marine surfaces of tidal erosion formed during the first period of falling stage shoreline regression. Shale capping this interval formed during delta abandonment and transgression. Continued sea level fall and delta progradation produced the second interval of tidal sandstone overlying a surface of marine tidal erosion. High-relief erosion surfaces cut into the second interval of tidal sandstone record the enlargement of distributaries by tidal currents after abandonment. A shell lag at the top of this interval of sandstone records transgressive ravinement. Tidal flat deposits above this surface aggraded as shoreline deposits vertically aggraded seaward of the areas studied. Minor meandering channel deposits occur within these tidal flat deposits.

Progradational intervals are capped by regionally traceable ferroan-dolomite cemented horizons. Discontinuous, patchy cements are found in regressive distal-shoreface sediments. Sandy sediments less than five meters below coal or carbonaceous shale tend to be leached greenish-white by the downward flux of acidic fluids during lowstands.

Although the upper Sego Sandstone is interpreted to have formed in a regressive tide-dominated setting, flood-oriented sedimentary structures dominate tidally influenced bedsets. This reflects preferential preservation of bedload transported by flood tides accelerated onto the delta front, rather than ebb-oriented currents carried basinward in a hypopycnal plume. On analogue deltas a salt wedge tended to lift hypopycnal ebb flows and enhances bedform migration under flood-tidal currents. Delta progradation was maintained by fall out of ebb directed sediments suspended above the sediment bed within hypopycnal flows, by ejection of the salt wedge by freshet flows, and by the production of ebb moving fluidized mud during major river floods.

Deep incisions into the upper Sego Sandstone are related to scouring of channels after distributary abandonment. After scouring, these narrow and deep distributaries were filled during renewed delta progradation or the transgressive reworking of sediment along the coast. Channels filled with thinly-bedded heterolithic deposits were preserved where distributaries were never reoccupied or were filled with estuarine sediments. This mechanism for the formation of high-relief erosion surfaces cut into a delta front does not require changes in sea-level and can occur during any stage of sea level.

Sequence stratigraphic interpretation of the upper Sego Sandstone is problematic because of differing opinions regarding the significance and interpretation of abrupt facies changes. The two intervals of sandstone most likely record falling stage systems tracts. High-relief incisions in the upper interval may not correlate to a lowstand surface of erosion observed west of the survey area, and thus may not be deposits of a late lowstand systems tract. Alternatively they are interpreted to have formed by the scouring and filling of distributary channels during highstand progradation.

The lack of studies documenting bedform migration on tide-influenced river deltas makes it difficult to find modern analogues for the upper Sego Sandstone. The Fly River delta in Papua New Guinea is a possible analog. Processes interpreted from the upper Sego Sandstone that resemble those documented to occur on the Fly River delta include distributaries enlarged by tidal currents, preferential preservation of landward-transported sediments, and low preservation potential of flooding events.

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